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HYDROGEOLOGY AND HYDROCHEMISTRY OF SPRINGS IN MANTUA VALLEY

AND VICINITY, NORTH-CENTRAL UTAH

by

Karen C. Rice

A thesis submitted in partial fulfillment
of the requirements for the degree

of

MASTER OF SCIENCE

in

Geology

Approved:

UTAH STATE UNIVERSITY
Logan, Utah

1987

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Karen C. Rice

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ABSTRACT

Hydrogeology and Hydrochemistry of Springs in Mantua Valley

and Vicinity, North-Central Utah

by

Karen C. Rice, Master of Science

Utah State University, 1987

Major Professor: Dr. James McCalpin
Department: Geology

Chemical and tritium analyses of groundwater, precipitation and discharge records, fracture orientations, lineaments, and structural, stratigraphic, and topographic relationships have been used to describe the groundwater systems of Mantua Valley, north-central Utah. Groundwater flows through fractured Paleozoic quartzites and carbonate rocks and discharges from eleven perennial springs in Mantua Valley. Permeability in quartzites is the result of intense faulting and jointing. Groundwater in carbonate aquifers flows through fractures and/or fractures modified by solution and discharges as relatively large springs (up to 227 liters per second). Neogene normal faulting, rather than extensive karst processes, has produced valleys which are closed or nearly closed to surface-water drainage. Groundwater in the area has relatively low total dissolved solids, is warmer than the mean annual air temperature, and is of the calcium-magnesium-bicarbonate type. Temperatures of the groundwater suggest circulation depths in excess of 10 to 185 meters. Intermittent turbidity and fluctuations in calcite and dolomite saturation indices and in groundwater temperatures suggest that springs may be supplied by mixtures of shallow and deeper groundwater flow. With the methods used here, a water budget analysis of the area indicates that recharge to the groundwater systems is approximately 49% of mean annual precipitation. Annual recharge and average discharge of the springs were used to calculate recharge areas, which range from 3.0 km^2 to 18 km^2. Tritium analyses of two of the springs suggest mean residence times of less than ten years.

(123 pages)

INTRODUCTION

A variety of geologic processes and climatic regimes affecting a geologic terrane through time can create complex conditions which control groundwater flow. Especially complex groundwater flow systems in carbonate rocks often result from karst activity, which can affect soluble formations to various degrees. Hydrogeologic studies which identify geologic conditions and the resulting controls on groundwater movement can lead to the characterization of groundwater systems.

Fifty percent of the U.S. population depends on groundwater for its primary source of drinking water (U.S. Library of Congress, 1985; Gass, 1986a), and many aquifers are rapidly becoming polluted or depleted (U.S. Library of Congress, 1985; Gass, 1986b). Because of the growing awareness and concern over groundwater resources, hydrogeologic studies are becoming increasingly necessary for managing groundwater supplies. Understanding groundwater systems is essential to adequate management of valuable groundwater resources. The purpose of this study is to obtain an understanding of the hydrogeology of Mantua Valley, Utah, and to characterize the hydrogeologic regimes which exist in the vicinity.

Location and General Features

The study area is located in north-central Utah, on portions of the Mount Pisgah, Mantua, and James Peak, Utah 1:24,000 topographic quadrangles (fig. 1). The study area comprises approximately 104 km^2 (25,600 ac), of which about three-fourths is mountainous and about one-fourth is valley terrain. Mantua Valley, situated near the center of the study area, is approximately 40 km (25 mi) southwest of Logan and 6

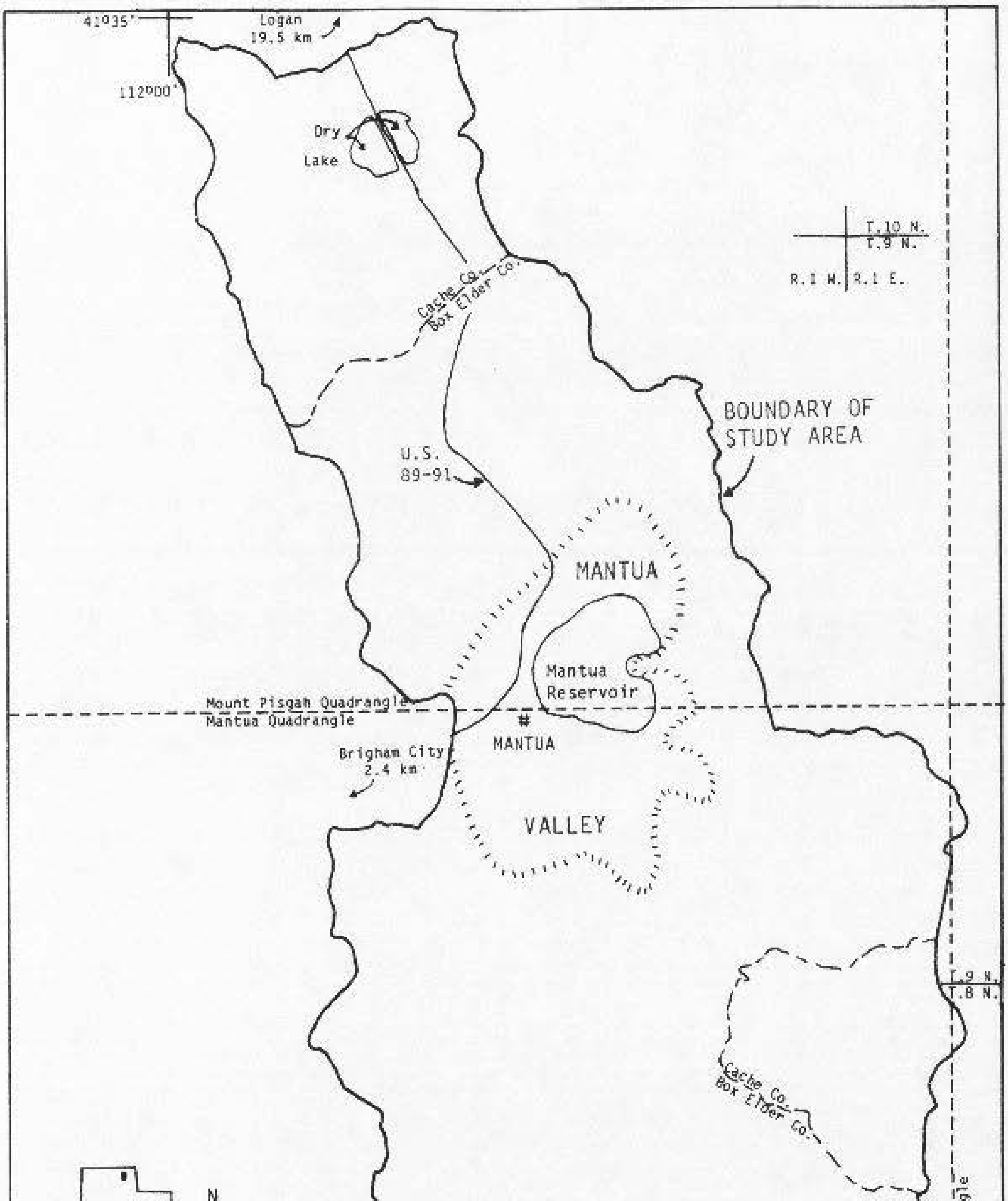


Figure 1.--Location of study area.

km (3.5 mi) east of Brigham City. The study area is about 4 km (2.5 mi) wide at the northern boundary and widens southward to about 9 km (5.5 mi). The north-south length of the study area is about 18.5 km (11.5 mi).

Major valleys located within the study area (pl. 1), from north to south, and their elevations, are as follows: Dry Lake (1731 m, 5680 ft), Mantua (1585 m, 5200 ft), Clay (1853 m, 6080 ft), Sink Hole (2001 m, 6565 ft), and Devils Gate (1975 m, 6480 ft). Dry Lake and Sink Hole Valleys are closed to surface-water drainage, while Mantua, Clay, and Devils Gate Valleys are in the Box Elder Creek drainage basin (pl. 1).

The northwestern part of the study area is bounded by the Wellsville Mountains, which have elevations in the study area up to 2390 m (7841 ft). The Pisgah Hills, separated from, but continuing the general trend of the Wellsville Mountains (Williams, 1948), occupy the northeastern portion of the area. The Wasatch Mountains, with elevations up to 2502 m (8208 ft) in the study area, and the South Hills (Williams, 1948) dominate the southern portion of the study area.

Climatological data are collected by the National Oceanic and Atmospheric Administration at the Hardware Ranch and the Huntsville Monastery weather stations. Data from these stations were considered to be more representative of climatic conditions in the study area than the Brigham City weather station, because the Hardware Ranch and Huntsville Monastery stations are located at elevations similar to those in the study area. The study area is characterized by low to moderate precipitation, moderately cold winters, warm dry summers, and large daily temperature changes. The mean annual temperature is approximately 6^{o} C

(43^{o} F), (10 year record, National Oceanic and Atmospheric Administration, 1976-85). The area normally receives from 635 to 762 mm (25-30 in.) of precipitation annually (29 year record, 1931-1960; Bjorklund and McGreevy, 1974, fig. 11, p. 47). Most of the precipitation falls as snow, which covered the valley floors during parts of November and December, 1985, and January, February, and March, 1986.

Much of the land in the study area is part of the Cache National Forest, although some of it is state and privately owned. The land within the study area is used primarily for agriculture and grazing; other uses include raising cutthroat trout and recreation. Native vegetation is predominantly grasses, big sagebrush (Artemisia tridentata), and Rocky Mountain Maple (Acer grandidentatum).

Approach to the Problem

Objectives of the study are to: 1) identify hydrostratigraphic units; 2) determine the origins of anomalous topographic features in the study area such as large-scale closed and nearly closed depressions; 3) estimate the primary and secondary porosities of formations in the study area; 4) identify structural and stratigraphic controls on groundwater movement; and 5) develop conceptual flow system models which describe depths of groundwater circulation, groundwater flow paths, and recharge areas for five springs in the study area, and which estimate mean residence times for two of the springs.

By collecting and integrating field and laboratory data to resolve the objectives of the study, the groundwater flow systems in the vicinity of Mantua Valley can be described. The field analysis can provide an overview of the geologic setting, including the intrinsic controls which stratigraphy, structure, and topography have on groundwater movement. Field data required include: 1) extent of solution which has occurred along bedding planes and joints in carbonate rocks; 2) orientations of bedding planes, joints, and lineaments; 3) apertures and frequencies of fractures in rock outcrops; and 4) primary porosities of formations in the study area. In addition, analysis of groundwater samples will provide data on the temperature, pH, alkalinity, and the hydrochemistry of groundwater in the study area. The laboratory analysis can provide a characterization of the chemical nature of groundwater in Mantua Valley. Laboratory data required include: 1) major inorganic ion concentrations of groundwater samples; 2) tritium activities of groundwater from two of the springs; 3) calculations of saturation indices from major ion concentrations; and 4) X-ray diffraction patterns of filtrates from turbid water samples.

Previous Investigations

Mantua Valley Study Area

The Paleozoic rocks of the Logan, Utah 30' quadrangle have been mapped by Williams (1948) at a scale of 1:125,000. The 30' quadrangle includes the northern half of the Mantua study area. A 1:24,000 scale geologic map produced by Gelnett (1958) covers the study area west of U.S. Route 89-91. The southern half of the Mantua study area is covered by Crittenden and Sorensen's (1985) 1:24,000 scale geologic map of the Mantua quadrangle and part of the Willard quadrangle. Ezell (1953) produced a 1:21,120 scale geologic map of the southeastern portion of the study area.

Several geologic maps, covering the Mantua Valley study area, have been compiled from previously published and unpublished works. Stokes (1963) compiled a 1:250,000 scale map of northwestern Utah. Doelling and others (1980) prepared a 1:125,000 scale geologic map of Box Elder County. A 1:100,000 scale geologic map of the Northern Wasatch Front was compiled by Davis (1985). Dover (1985) compiled a 1:100,000 scale geologic map of the Logan 30' X 60' quadrangle.

Bjorklund and McGreevy (1974) evaluated the groundwater resources of the lower Bear River drainage basin in Box Elder County. That report includes information on the geology, discharge, and chemical quality of groundwater in parts of Box Elder County, with a subsection on Mantua Valley.

Price and Jensen (1982) investigated the surface-water resources of the Northern Wasatch Front. Their 1:100,000 scale map covers the Mantua Valley study area and includes information on runoff and dissolved-solids concentrations of rivers along the Northern Wasatch Front.

Regional Hydrostratigraphic Units

Bjorklund and McGreevy (1971) and Davis (1985) described in general terms the water-bearing properties of many of the stratigraphic units which are exposed in the study area. The Precambrian units are described as aquitards because of their low permeabilities, which are the result of cementation, compaction, or crystallization (Bjorklund and McGreevy, 1971); however, they and Davis (1985) reported that the Precambrian units are locally permeable and may provide a large volume for groundwater where they are fractured and faulted. Bjorklund and McGreevy

(1971) suggested that permeability of the Paleozoic units is the result of fractures and solution along fractures and bedding planes, and that all the formations probably yield water to at least a few small springs. Davis (1985) reported that most of the Paleozoic units are probably aquifers where groundwater is available. Bjorklund and McGreevy (1971) suggested that the Wasatch and Evanston Formations have low permeabilities, with some moderately permeable beds. Bjorklund and McGreevy (1974) reported that landslide, colluvium, and slopewash deposits generally have low to high permeabilities, and alluvial deposits have moderate to high permeabilities.

GEOLOGIC SETTING

The study area is situated within the Wasatch Range subdivision of the Middle Rocky Mountains physiographic province (Stokes, 1977, fig. 1). The Wasatch Range extends southward from the Bear River Narrows and merges into the Colorado Plateau south of Nemo, Utah. The Wasatch fault zone defines the western edge of the range, while the eastern boundary is marked by the break between high slopes of the range and the Wasatch Plateau (Stokes, 1977). Stokes (1977) subdivided the range into three segments; the study area lies within the Northern Wasatch segment, defined as the area north of the Weber River.

Stratigraphy

Carbonate and quartzite rock units and unconsolidated deposits characterize the stratigraphy of the study area. Rock units range in age from Precambrian through Mississippian, while unconsolidated to semiconsolidated deposits are Tertiary and Quaternary in age. Pennsylvanian and Permian units are not exposed within the study area, while Mesozoic units are absent due to a regional unconformity. Figure 2 shows a stratigraphic column of units exposed in the study area with thicknesses and brief descriptions of the units. The following descriptions and thicknesses of stratigraphic units have been reported by Crittenden and Sorensen (1985).

Precambrian Units

The Precambrian section in the study area, described by Crittenden and Sorensen (1985), is characterized by vitreous quartzites, sandstones,

Age	Formation and Description	Lith-ology	Thick-ness (m)
Quaternary	Unconsolidated Deposits - colluvium, slope-wash, alluvium, terrace deposits		0-154+
Tertiary	Norwood Tuff - tuffaceous silt, sandy tuff		0.5
	Wasatch and Evanston Formations - quartzite boulders, weakly cemented		0-75
Mississippian	Brazer Formation - limestone, argillaceous limestone, cherty limestone, silty lime-stone, shale		844
	Humbug Formation and Deseret Limestone - quartzitic sandstone, limestone, dolostone, chert, phosphate shale		800
	Lodgepole Limestone - limestone, chert		230-275
Devonian	Hyrum (?) Dolostone - sandy dolostone, chert		15-50
	Water Canyon (?) Formation - silty dolostone		15-50
Silurian	Laketown Dolostone - dolostone, chert		460
Ordovician	Fish Haven Dolostone - dolostone, chert		25-50
	Swan Peak Formation - shale, quartzite		10-75
	Garden City Formation - limestone, silty limestone, dolostone, chert, intraforma-tional conglomerate		250-370
Cambrian	Nounan and St. Charles Dolostones - dolo-stone, limestone, quartzite, quartz sandstone		580
	Bloomington Formation - siltstone, mudstone, shale, limestone, silty limestone, dolo-stone		405-545
	Blacksmith Dolostone - dolostone, limestone, silt horizons		210-245
	Langston Dolostone and Ute Limestone - lime-stone, dolostone, siltstone, shale		150-245
	Geertson Canyon Quartzite - quartzite		1200
Precambrian	Browns Hole Formation - vitreous quartzite		35-85
	Mutual Formation - quartzite		675-800
	Inkom Formation - sandstone, siltstone, tuff		5-60
	Caddy Canyon Quartzite - vitreous quartzite		300-500
	Papoose Creek Formation - siltstone, quartz-itic sandstone		230-460

Figure 2.--Stratigraphic column of units in the study area (compiled from Crittenden and Sorensen, 1985 and Dover, 1985).

and siltstones of upper Proterozoic age. The thickness of the Precambrian section ranges from 1245 m to 1905 m. The oldest formation in the area, Papoose Creek, is overlain by rocks of the Brigham Group. Formations, from oldest to youngest, composing the Brigham Group in the study area are named Caddy Canyon, Inkom, Mutual, and Browns Hole (fig. 2).

Paleozoic System

The Paleozoic system in the study area, described by Crittenden and Sorensen (1985), is dominated by carbonate rocks and includes two quartzite units (Cambrian Geertson Canyon Quartzite and the Ordovician Swan Peak Formation) and a minor quartzitic sandstone unit (Mississippian Humbug Formation). Total thickness of the Paleozoic section ranges from 5179 m to 5739 m. Two disconformable contacts exist within the section, one between the Middle Ordovician Swan Peak Formation and the Upper Ordovician Fish Haven Dolostone, and the second between the Fish Haven and the Middle and Upper Silurian Laketown Dolostone.

The base of the Paleozoic section contains the upper member of the Brigham Group, the Geertson Canyon Quartzite, of Lower Cambrian age. The 1200 m section of Geertson Canyon Quartzite includes a basal arkosic conglomerate, medium- to coarse-grained quartzites, lenses of small pebbles, and *Skolithos* tubes at the top. The Middle Cambrian age Langston Dolostone and Ute Limestone, with a combined thickness of 150-245 m, overlie the Geertson Canyon Quartzite. Coarse-crystalline Langston dolostone beds lie below medium- to thin-bedded, fine- to medium-crystalline Ute Limestone beds, which are interbedded with shale and siltstone. The 210-245 m thick section of Blacksmith Dolostone, of Middle Cambrian age, consists of medium- to thin-bedded, fine- to medium-crystalline dolostone and limestone, with local silt horizons. The overlying Middle Cambrian age Bloomington Formation is divided into a limestone and two shale members. The Hodges Shale Member is 135-145 m of siltstone, mudstone, and shale. The overlying Unnamed Limestone Member consists of fine- to medium-crystalline limestone and silty limestone with one or more dolostone beds, and is 150-215 m thick. The Calls Fort Shale Member lies at the top of the Bloomington Formation and is 120-185 m of interbedded shale and limestone. The Middle and Upper Cambrian Nounan Dolostone and the Upper Cambrian St. Charles Dolostone have a combined thickness of 580 m and consist of coarse- to medium-crystalline dolostone and limestone. The base of the St. Charles Dolostone is marked by the Worm Creek Quartzite Member, composed of quartzite and quartz sandstone (Dover, 1985).

The overlying Lower and Middle Ordovician Garden City Formation, 250-370 m thick, is composed of coarse- to medium-crystalline limestone, bioclastic grains, silty limestone and dolostone, chert, and intraformational conglomerate. The Middle Ordovician Swan Peak Formation contains 10-75 m of shale and quartzite with fucoidal markings. The 25-50 m thick Upper Ordovician Fish Haven Dolostone unconformably overlies the Swan Peak and is characterized by coarse- to medium-crystalline dolostone, with bioclastic grains and chert.

The Silurian period is represented by only one unit, the Middle and Upper Silurian Laketown Dolostone. The approximately 460 m thick section of Laketown Dolostone consists of very coarse- to medium-crystalline cliff-forming dolostone, with scattered lenses of chert.

The Lower Devonian Water Canyon Formation is composed of 15-50 m of fine-crystalline, thin-bedded to laminated, silty dolostone. The overlying Middle and Upper Devonian, 15-50 m thick, Hyrum Dolostone is characterized by coarse-crystalline to sandy dolostone, with lenses of chert.

Fine- to medium-crystalline, fossiliferous limestone with chert makes up the 230-275 m thick section of Lower Mississippian Lodgepole Limestone. The overlying Lower and Upper Mississippian Deseret Limestone and the Upper Mississippian Humbug Formation have a combined thickness of approximately 800 m. The Deseret Limestone, correlative to Williams' (1948) lowest portion of the Brazer Formation, comprises limestone, dolostone, chert, and occasional exposures of phosphate shale at the base of the unit. The Humbug Formation is medium- to coarse-grained quartzitic sandstone, with interlayered limestone and dolostone, and is correlative to Williams' (1948) Brazer Formation. The top of the Paleozoic section is marked by the Upper Mississippian Brazer Formation (Williams, 1948), correlative with the Great Blue Limestone and the Manning Canyon Shale (Hintze, 1973, p. 39). The 844 m thick section of Brazer Formation at Dry Lake comprises limestone, argillaceous limestone, cherty limestone, silty limestone, and black shale (Williams, 1948).

Tertiary System

The Tertiary system, described by Crittenden and Sorensen (1985), is represented in the study area by only two units, which are separated from the Paleozoic section and from each other by unconformities. The Wasatch and Evanston Formations of Eocene, Paleocene, and possibly Upper Cretaceous age are unconsolidated units with a combined thickness of 0-

75 m. They are characterized by reddish-brown weathering cobbles and boulders derived from the Geertson Canyon Quartzite and late Precambrian quartzites (Crittenden and Sorensen, 1985). The Lower Oligocene and Upper Eocene Norwood Tuff has very limited exposure in the study area and attains a thickness of only 0.5 m. The Norwood Tuff is an extensively altered, thin-bedded tuff, tuffaceous silt, and sandy tuff.

Quaternary System

Unconsolidated Quaternary units in the study area are generally found in valleys and along streams. Crittenden and Sorensen (1985) divided the alluvium of Devils Gate Valley into two members: 1) the 0-38 m thick lower member, a tuffaceous clay and silt that was probably derived from the Norwood Tuff; and 2) the 0-35 m thick upper member, consisting of red to brown sandy clay and silt with lag gravels, which was probably derived from the Wasatch Formation. Crittenden and Sorensen (1985) have also mapped Holocene and Pleistocene landslide deposits, Holocene and Pleistocene slopewash deposits and colluvium, and Holocene alluvial deposits in the study area.

Structure

Because Williams (1948) mapped northeast-trending tear faults in the northern part of the study area, which were not located during field analysis for this study, Dover's (1985) mapping was used for structural interpretations. Structural features of the study area include joints, north-, northeast-, and northwest-striking normal faults, the Box Elder thrust fault, and two other minor thrust faults. All exposed Paleozoic stratigraphic units in the study area are part of the upper plate of the eastward-dipping Willard thrust fault, located southwest of the study area (Crittenden and Sorensen, 1985). Strikes of stratigraphic units in the southern part of the study area vary from nearly east-west to north-south and average N. 50° W., with an average dip of 40° NE. Strikes of stratigraphic units in the northern part of the study area are more consistently oriented, with an average strike and dip of N. 35° W., 60° NE.

From the beginning to the end of the Cretaceous period, the study area underwent a compressional phase, termed the Sevier orogeny by Armstrong (1968). This compressional phase is evidenced in the region by eastward-directed overthrusts, with displacements of tens of miles, and several large, compressional folds (Armstrong, 1968). The Laramide orogeny coincided with and closely followed the Sevier orogeny and is characterized by vertical uplifts which occurred from latest Cretaceous time through the Eocene epoch (Grose, 1972). The ancestral Wasatch Mountains were elevated by Laramide uplifts, and became the source of the Tertiary-age Wasatch Conglomerate. The essentially horizontal Wasatch Conglomerate mantled all older folded strata during the Eocene epoch (Hintze, 1973, p. 74). At present, it is an unconformable erosional remnant, recognized as rounded quartzite boulders in a fine-grained red matrix, mantling hilltops.

Following compression of the Sevier-Laramide orogeny, the area experienced a change in tectonic forces, which resulted in Basin and Range extension. Stewart (1971) determined that Basin and Range extension was initiated 17 million years ago and peaked during the last 7 to 11 million years. Historic seismic evidence (Cluff and others, 1970) and alluvial fans younger than Lake Bonneville which are cut by fault scarps (Personius, 1986) indicate that the region has experienced extensional deformation through the present. Typical Basin and Range structure is expressed as north-trending basins and ranges bounded by north-striking faults (Stewart, 1971).

Hydrology and Hydrogeology

Mantua Valley Study Area

The study area is situated in the Western Mountain Ranges groundwater region of the United States which is characterized by mountains alternating with narrow alluvial valleys (Heath, 1984). Slopes and summits of mountains consist of bedrock exposures or boulders and thin soils covering bedrock; valleys are filled with coarse alluvium transported from the surrounding mountains (Heath, 1984). He further reported that within the Western Mountain Ranges, groundwater recharge is from rainfall and from snowmelt at the higher elevations.

The study area forms the southeastern portion of the lower Bear River drainage basin and is drained by Box Elder Creek westward to Great Salt Lake. Box Elder Creek originates in Devils Gate Valley and flows northward toward Mantua Valley over both bedrock and alluvium. For a four-year period of record (1959-63), Box Elder Creek, gauged near the NW corner sec. 34, T. 9 N., R. 1 W., had an average discharge of 53.8 l/s (853 gpm) and ranged from 0.0 to 8212 l/s (0.0-130,161 gpm; Price and Jensen, 1982).

Mantua Valley is a groundwater discharge area for the surrounding mountain and valley terrain (Bjorklund and McGreevy, 1974). The topography of Mantua Valley and locations of springs are shown on figure 3. Eleven perennial springs discharge at rates of 0.6 to 271 l/s (10-4300 gpm) from the perimeter of Mantua Valley. Some of the spring water is piped directly to an aqueduct through which it is transported by gravity flow to Brigham City for municipal use. The rest of the spring water discharges into Mantua Reservoir, which is used for recreation as well as for storage of Brigham City's irrigation-water supply. The municipal water supply for the town of Mantua is supplied by two springs discharging from the extreme southwestern end of Mantua Valley (named "Mantua Spring" for this report) and from a well in Mantua Valley.

The geology and uses of the springs analyzed for the study are summarized in table 1. Peter Jensen, West Hallings, Mud, and Maple Springs all issue from carbonate rocks or from alluvium overlying carbonate rocks and are collectively referred to as "the carbonate springs." Mantua Spring discharges from quartzites and is referred to as "the quartzite spring."

Previous Investigations in Carbonate Hydrogeology

White (1969) developed conceptual models characterizing carbonate aquifers, based on their flow characteristics. The classification divides carbonate aquifers into three main types: 1) diffuse flow aquifers; 2) free flow aquifers; and 3) confined flow aquifers. Because interpretations of some of the springs in this study are based on White's (1969) conceptual models of carbonate aquifers, two of the models are described in detail below.

Diffuse flow aquifers are characterized by carbonate rocks that

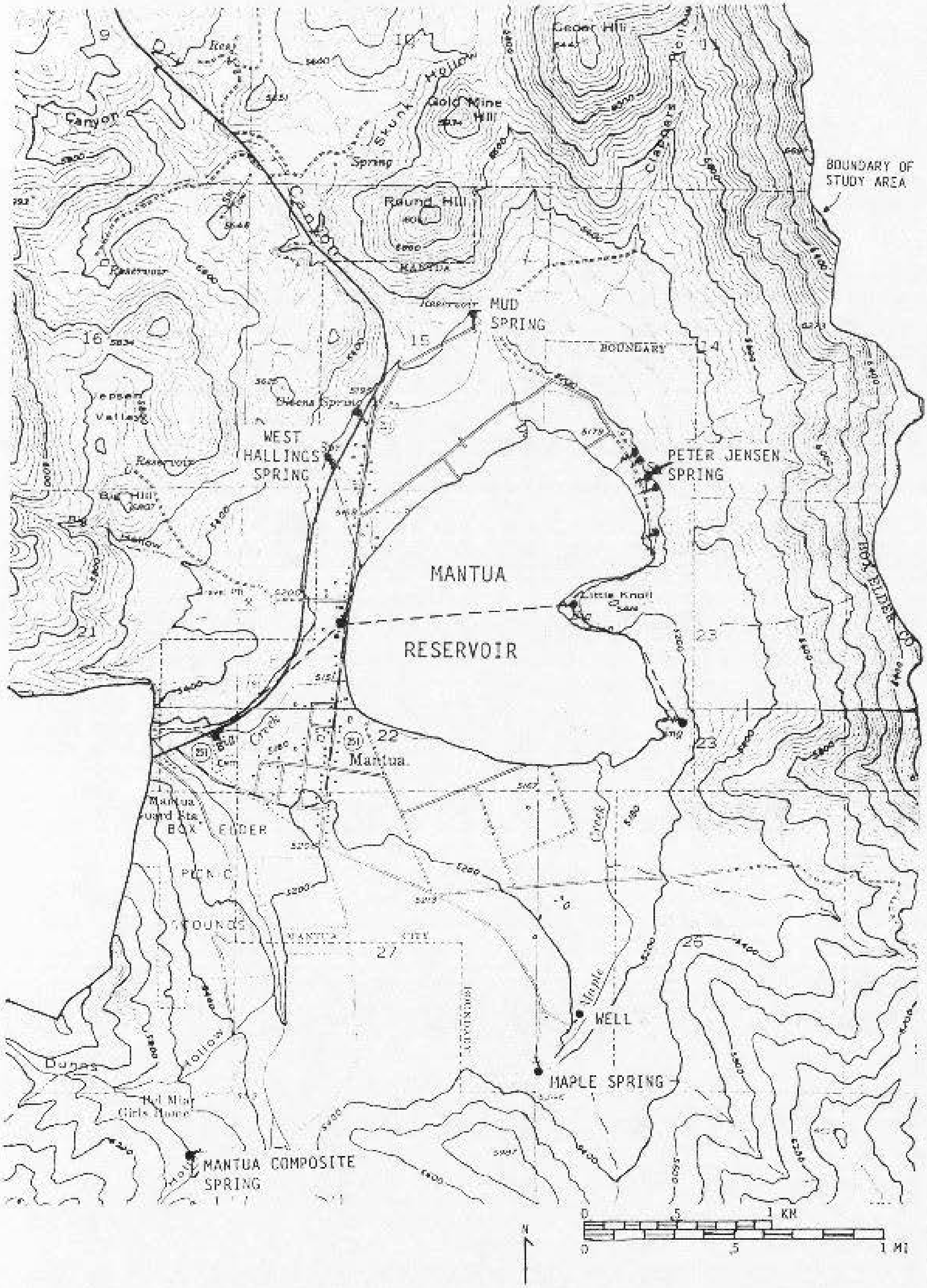


Figure 3.--Topographic map of Mantua Valley showing locations of springs. Dashed lines show collection system for Brigham City municipal water supply. Contours in feet; 1 ft = 0.3048 m.

Table 1.--Summary of sampled spring locations and uses (n.d., no data)

Station	Geology at Spring Orifice[a]	Topographic Situation	Elevation (m) (ft)	Use	Discharge[b] (l/s; gpm) (date)
Peter Jensen Spring	Quaternary alluvium overlying Silurian (?) carbonate rocks	NE edge of Mantua Valley	1585 5200	Portion of Brigham City municipal water supply, contributes to Mantua Reservoir	31; 491 (1924)
West Hallings Spring	Quaternary alluvium overlying Cambrian Bloomington Formation	Western edge of Mantua Valley	1579 5180	Portion of Brigham City municipal water supply, contributes to Mantua Reservoir	227; 3590 (1958)
Mud Spring	Quaternary alluvium overlying Ordovician Garden City Formation	Area of springs and seeps at base of Round Hill, north end of Mantua Valley	1585 5200	Supplies Mantua irrigation ditches, contributes to Mantua Reservoir	99; 1569 (9-28-70)
Mantua Spring (composite)	Cambrian Geertson Canyon Quartzite	Along bedrock wall in extreme SW Mantua Valley	1768 5800; 1890 6200	Portion of Mantua municipal water supply	3; 45 (10-71)
Maple Spring (Upper)	Calls Fort Shale Member of Cambrian Bloomington Formation	Base of hill at contact with terrace deposits, S end of Mantua Valley, one source of Maple Creek	1591 5220	Mantua Fish Hatchery, supplies irrigation ditches, contributes to Mantua Reservoir	90; 1433 (9-28-70)
Well	Pleistocene terrace deposits (?)	In alluvium of SE Mantua Valley	1583 5195	Domestic and irrigation	n.d.[c]

[a] from Dover (1985), and Crittenden and Sorensen (1985)

[b] from Bjorklund and McGreevy (1973); may not be representative of yearly discharge; Upper Maple Spring after oral communication with Ron Roubidoux, 1985

[c] unknown; other wells in Mantua Valley discharge water at rates of 16-120 l/s (260-1900 gpm); and at least two wells are under artesian conditions

have undergone little modification by solution. Modification by solution is expressed as widened joints or bedding planes, which are well-connected, and karst topography is subdued. Coarse-crystalline dolostones and shaley limestones usually constitute diffuse flow aquifers, because such lithologies are less susceptible to solution than pure limestones. Discharge from diffuse flow aquifers is through many small springs and seeps, and the water table is usually well-defined.

Free flow aquifers (conduit aquifers of Shuster and White, 1971) have undergone extensive solution so that groundwater flows through highly connected systems of conduits. Rocks surrounding the conduits typically have low porosities and permeabilities. Free flow aquifers are overlain by well-developed karst topography. Groundwater flows frequently reach the turbulent regime with velocities of tens of feet/second. Main conduits of flow transport a bedload as well as a suspended load of sediment and may be considered as underground continuations of surface streams. Groundwater in drainage basins of tens or hundreds of square miles is discharged through a single, large spring.

White (1979), in a study of karst landforms in the Wasatch Mountains, found that the karst developed on extensive outcrop areas of Mississippian carbonate rocks is sparse and subdued. Karst features such as solution along joints and fractures are numerous, but dolines are absent and internal drainage is rare (White, 1979). He reported that caves in the Wasatch Mountains are sparsely distributed, small, and of only local extent [eight caves, all less than 914 m (3000 ft) in length]. White (1979) concluded that karst development has been restricted because of thick sections of dolostone and dolomitic limestone, which do not dissolve as readily as limestones.

Karstification of some of the Paleozoic carbonate formations exposed in the study area is known to have occurred in the Bear River Range and in the northern Wasatch Front. Davis (1985) stated that the Bloomington Formation has produced karst topography and that the Garden City Formation and the underlying formations have been partially dissolved to form Devils Gate Valley. Bjorklund and McGreevy (1971) reported that large caves, developed by solution, are present in the Garden City Formation and the Lodgepole Limestone in the Bear River Range. They also found that sinkholes have developed in the Bloomington and Garden City Formations and the Laketown and Hyrum Dolostones from solution and collapse of the overlying rock, and that solution channels in these carbonate rocks permit large springs to issue from them. Bjorklund and McGreevy (1971) also reported that springs discharging more than 32 l/s (500 gpm) flow from the Langston, Blacksmith, Bloomington, Nounan, St. Charles, Garden City, Laketown, Hyrum, and Lodgepole Formations.

DATA COLLECTION

Field Analysis

Field data were collected as an aid to the characterization of groundwater flow systems in the vicinity of Mantua Valley. Four types of field data were obtained: 1) data on the orientations of joints, bedding planes, and faults for fracture trace analysis; 2) data on the apertures and density of joints and bedding planes for estimates of secondary porosity; 3) data on the primary porosity of non-carbonate formations; and 4) groundwater samples for hydrochemical analysis and turbidity data.

Joint, bedding plane, and fault orientation measurements were taken during the summer months of 1985. Rock outcrops and roadcuts were analyzed with Brunton pocket transit, tape measure, dilute HCl acid (10%), and hand lens. Strikes and dips of bedding planes and all joints were recorded at 96 outcrops, and dominant joint orientations were identified and recorded.

Data for estimates of secondary porosity of eleven formations in the study area were collected following the procedure outlined by Bianchi and Snow (1969). They recommended a random sampling of fracture orientations, because it is usually not possible to measure the orientation of every fracture. All joints which intersect a sampling line of known length, oriented perpendicularly to the majority of the fractures, are measured (Bianchi and Snow, 1969). They further reported that because a large number of orientations are measured and the sampling localities are randomly chosen, it is assumed that fracture spacings and widths not measured are identical to those measured along the sampling line. Secondary porosities of outcrops in the study area were estimated by multiplying the average width of the observed fractures in an outcrop by the total number of fractures, dividing by the length of the sampling line, and multiplying by 100. Estimates made this way are probably maximum values of secondary porosity, since fractures will decrease in width with depth and some will disappear. In addition, errors in sampling, such as not having a sufficient length of sample line to be representative of the outcrop, may affect the estimates. Outcrops adjacent to identified faults were observed for any increase in fracture spacing.

Primary porosity of non-carbonate rock samples was determined with a simple field test developed by Robert Q. Oaks, Jr. (Utah State University, oral commun., 1984). The amount of time required for one drop of 10% HCl acid, placed on a fresh surface of the sample, to soak in is recorded. The porosity of the sample then can be approximated from table 2.

Black and white, 1:20,000 scale aerial photographs of the area were examined and used as an aid to field work. They were also used to help identify lineaments, sinkholes, areas of high groundwater, joints, and specific geologic units.

Groundwater data were collected on a monthly to bimonthly basis starting March 1985 and ending May 1986. Of the eight springs initially sampled, five springs and a well were chosen for continued sampling, based on the desire to obtain a variety of magnitudes of discharge, locations within Mantua Valley, chemistries, and lithologies at the point of discharge. Peter Jensen and West Hallings Springs were sampled

Table 2.--Primary porosity determination of non-carbonate rock samples

Time (seconds)	Porosity (%)	Description
>15	trace or none	very poor
10-15	1-5	poor
5-9	6-11	moderate
0-4	12-15	good

from concrete spring boxes. Mud and Maple Springs were sampled at their natural outlets. Maple Spring is a combination of two springs - an upper spring which discharges from bedrock, and a lower spring about 31 m (100 ft) away which upwells as springs and seeps through alluvium. Only Upper Maple Spring was sampled and analyzed. Mantua Spring is actually a composite of two unnamed springs connected to a spring box which is gravity-fed to the town of Mantua. When access to the spring box was impossible, Mantua Spring was sampled from a fire hydrant connected to the system (7-17-85, 9-24-85, and 5-25-86), a house in the town of Mantua also connected to the system (1-19-86 and 3-15-86), or from an area of groundwater seepage topographically above the spring box (11-24-85). Analyses of Mantua Spring groundwater not sampled from the spring box or fire hydrant are reported in appendices A and C but were not used in any calculations involving temperature, ion concentrations, or saturation indices of Mantua Spring groundwater. Well-water was sampled from a faucet approximately 244 m (800 ft) from the well. A total of 55 samples was taken during the year.

At each sample site, non-quantitative data were recorded on relative discharge and turbidity of the springs. Groundwater samples were collected in one-liter polyethylene bottles that had been rinsed with 20% HNO_3 acid, then rinsed several times with deionized water. Two one-liter bottles of water were obtained at each sample site. Samples were taken as close to the spring openings as possible to avoid water that may have equilibrated with the atmospheric partial pressure of CO_2. Usually, water from the well was running prior to sampling, however when it was not, water was sampled after it ran for several minutes to avoid sampling water that had been standing in the casing. The samples were filtered through a 0.45 micron membrane filter to remove suspended solids, using a battery-powered GeoFilter Peristaltic Pump, Model #004, into clean one-liter bottles. One filtered sample from each site was stabilized by adding 50 ml of reagent-grade HNO_3 acid to obtain a pH less than 2.0, and the second filtered sample remained untreated. Acidified samples were stored in the laboratory at room temperature, while untreated samples were stored in a refrigerator at 4° C. Immediately after collection, the samples were analyzed for temperature, pH, and alkalinity. Temperature and pH were measured with a battery-powered Markson Digital pH/Temperature Meter, Model 90, which had been standardized in the lab. Metrepak buffers of pH 4, pH 7, and pH 10 were used as standards. Problems with instrumental drift were combatted by allowing the electrode several minutes of stabilization time in the sample before taking the pH reading. Alkalinity values were obtained with the low range method of the Hach Titration Method of Alkalinity (Alkalinity Test Kit, Model AL-AP). Alkalinity data were converted to grains per gallon as $CaCO_3$ by dividing the total number of drops of Sulfuric Acid Standard solution used by 2.5. The grains per gallon as $CaCO_3$ were converted to mg/l HCO_3^- with the relationship:

$$\text{grains per gallon } CaCO_3 \times \frac{17.1187 \text{ mg/l}}{1 \text{ grain/gal}} \times \frac{122.0122 \text{ g/2 moles } HCO_3^-}{100.0782 \text{ g/1 mole } CaCO_3}$$

Alkalinity of the samples is expressed as mg/l HCO_3^-.

Water samples for tritium analysis were collected from West Hallings and Maple Springs and from a snow avalanche approximately 2.0 km (1.25 mi) south of Black Mountain on 29 June 1986. Samples were collected in factory-fresh one-liter polyethylene bottles that had not been rinsed or allowed to overflow with sample water. Melted snow was quickly decanted into a polyethylene bottle, after it had been allowed to melt in an air-tight glass container. The samples were sent to the Tritium Lab at Rosenstiel School of Marine and Atmospheric Science in Miami, Florida for analysis.

Laboratory Analysis

Laboratory Methods

Fifty-five groundwater samples were analyzed for Na^+, K^+, Ca^{+2}, Mg^{+2}, and Cl^-, and forty-six samples were analyzed for SO_4^{-2}, SiO_2, and total dissolved solids (TDS). Samples to which HNO_3 acid was added in the field were used to analyze Na^+, K^+, Ca^{+2}, and Mg^{+2}, while untreated samples were used to obtain concentrations of SiO_2, SO_4^{-2}, Cl^-, and TDS. Laboratory methods of chemical analyses are summarized in table 3.

Concentrations of Na^+, K^+, Ca^{+2}, and Mg^{+2} were determined with a Perkin-Elmer Atomic Absorption Spectrophotometer, Model 303, using standard methods (Perkin-Elmer Corp., 1973). Samples were diluted according to the concentration of the ion in the sample and to the range

Table 3.--Summary of chemistry laboratory methods

Ion/Species	Concentrations of standards (mg/l)	Dilution of sample	Volume of sample (ml)	Method
Sodium	0.25, 0.50, 0.75, 1.00	1:10	10	Atomic Absorption
Potassium	0.5, 1.0, 1.5, 2.0	none	10	Atomic Absorption
Calcium	1.75, 3.50, 5.25, 7.00	1:10	10	Atomic Absorption
Magnesium	0.125, 0.250, 0.375, 0.500	1:100	10	Atomic Absorption
Chloride	1, 10, 100, 1000	none	50	Chloride Electrode
Sulfate	5, 10, 15	none	25	Visible Spectrophotometry
Bicarbonate[a]	none	none	15	Titration
Silica	5, 10, 15	none	10	Visible Spectrophotometry
TDS	none	none	50	Evaporation

[a]field analysis

in concentrations of the standards, determined by the upper detection limit of the Atomic Absorption Spectrophotometer. A set of four standards (mg/l) for each cation was prepared and analyzed. The concentrations of the standards and the corresponding absorbancies were used to develop linear regression curves, from which concentrations of each cation in the samples were obtained.

Chloride activities of samples were measured with an Orion Chloride Electrode, Model 94-17A and a Markson Digital pH/Temperature Meter, Model 90. A 1000 mg/l Cl^- solution was prepared by dissolving 1.65 g reagent-grade NaCl in one liter of deionized water. Standards with concentrations of 1, 10, and 100 mg/l Cl^- were prepared by serial dilution of the stock 1000 mg/l solution. An ionic strength adjustor was prepared and one ml was added to each 50 ml of sample and standard to maintain a constant background ionic strength during analysis. A linear regression curve was developed from the measured electrode potentials (mv) of the standards and log of concentrations of the standards (mg/l). The log of the concentration of Cl^- in the samples was then read from the linear regression curve.

Sulfate concentrations were obtained with the Turbidimetric method using SulfaVer 4 Sulfate Reagent for Water (Hach Chemical Company, 1975) and a Varian Series 634 UV-Visible Spectrophotometer. The spectrophotometer was set at a wavelength of 450 nm, visible light, double beam, with a slit setting of 1.0 nm, and was allowed to warm up for several hours. Standards with concentrations of 5, 10, and 15 mg/l SO_4^{-2} were prepared by dilution of a stock 2000 mg/l solution (prepared by dissolving 1.4797 g of dried, anhydrous Na_2SO_4 in 500 ml of 1% HCl). A reagent powder pillow, which causes a milky precipitate to form if SO_4^{-2} is present, was mixed with each 25 ml sample and standard; the mixtures were required to stand five minutes to allow the turbidity to fully develop. Samples and standards were analyzed in 10-mm plastic cells. For each analysis, an untreated sample or standard was loaded into both the reference and sample positions, and the spectrophotometer was adjusted to 100% transmission. The treated sample or standard was then loaded in the sample position, while the untreated water remained in the reference position, and the % transmission was recorded. Concentrations of sulfate in the samples were read from a linear regression curve developed from log % transmission of standards and the corresponding concentrations of the standards.

The Molybdate Blue Method for Silica (Rainwater and Thatcher, 1960) and a Varian Series 634 UV-Visible Spectrophotometer were used to determine concentrations of SiO_2 in the samples. The spectrophotometer was set at a wavelength of 700 nm, double beam, visible light, with a slit setting of 1.0 nm, and was allowed to warm up for several hours. Standards with concentrations of 5, 10, and 15 mg/l SiO_2 were prepared by dilution of a sodium silicate solution (1.00 ml = 0.050 mg SiO_2). Four reagents were prepared (0.25N hydrochloric acid, 5% ammonium molybdate, 1% Na_2EDTA, and 17% sodium sulfite) and were added to the standards, a blank, and 10 ml of each sample. After 30 minutes of stabilization time, the spectrophotometer was zeroed with the blank in the reference position and in one of the sample positions. While the blank remained in the reference position, the standards and samples were analyzed in 10-mm glass cells, and the absorbancies were recorded. Concentrations of SiO_2 in the samples were obtained from a linear regression curve developed from the concentrations of the standards and the corresponding absorbancies.

Total dissolved solids (TDS) concentrations in the samples were determined by weighing the residue on evaporation of filtered samples. Labeled beakers were dried in an oven at approximately 110° C for one hour, cooled in a $CaCl_2$ desiccator, then weighed. Fifty ml of sample were pipetted into each beaker and were allowed to evaporate on a hot plate under a fume hood. The beakers were then heated to approximately 110° C for one hour to drive off any water that was adsorbed onto the glass, cooled in the desiccator, and reweighed. The difference in weight is TDS expressed as grams TDS/50 ml water, which was converted to mg/l by multiplying by 20,000.

The computer program WATEQF, developed by Plummer and others (1976), was utilized to obtain saturation indices of eight minerals, including calcite, dolomite, and quartz. Input data for the WATEQF program are temperature (°C), pH, and concentrations (mg/l) of the following ions and species: Ca^{+2}, Mg^{+2}, Na^{+}, K^{+}, Cl^{-}, SO_4^{-2}, HCO_3^{-}, and SiO_2.

A Siemens X-ray diffraction unit was used to identify the filtrate from turbid groundwater samples. The samples were scanned from 2° 2θ to 35° 2θ at 2° 2θ per minute using Ni-filtered copper K_{α} radiation. Part of a clean filter was analyzed first to identify any "noise" that may have been introduced by the filter. The filters were mounted on glass slides with petroleum jelly. Two filters, one from Mantua Spring and one from Maple Spring, were analyzed.

The analyzed tritium data for Maple and West Hallings Springs were compared with atmospheric tritium data, taken at the Salt Lake City, Utah recording station, for the period 1963-1983. Yearly weighted means of tritium data (IAEA, 1981) were used for 1963-1969, and monthly tritium data were used for 1970-1971 (IAEA, 1975), 1972-1975 (IAEA, 1979), 1976-1979 (IAEA, 1983), and 1980-1983 (IAEA, 1986). The tritium data were weighted with yearly means of precipitation to compensate for unusually dry and wet months, i.e., if a particular month's precipitation is low relative to the year's mean, that month's tritium value is weighted less. The tritium data were weighted with the relationship:

$$\frac{\text{activity of sample X monthly precipitation}}{\text{monthly mean of the year's precipitation}} = \text{activity of sample weighted for precipitation}$$

The activity of each sample weighted for precipitation was then adjusted for radioactive decay with the equation: $A = A_0e^{-\lambda t}$, where A is the observed radioactivity, A_0 is the radioactivity at the time the water entered the aquifer, λ is the decay constant, and t is the age of the water.

RESULTS

Field Analysis

Hydrogen-Ion Activity (pH)

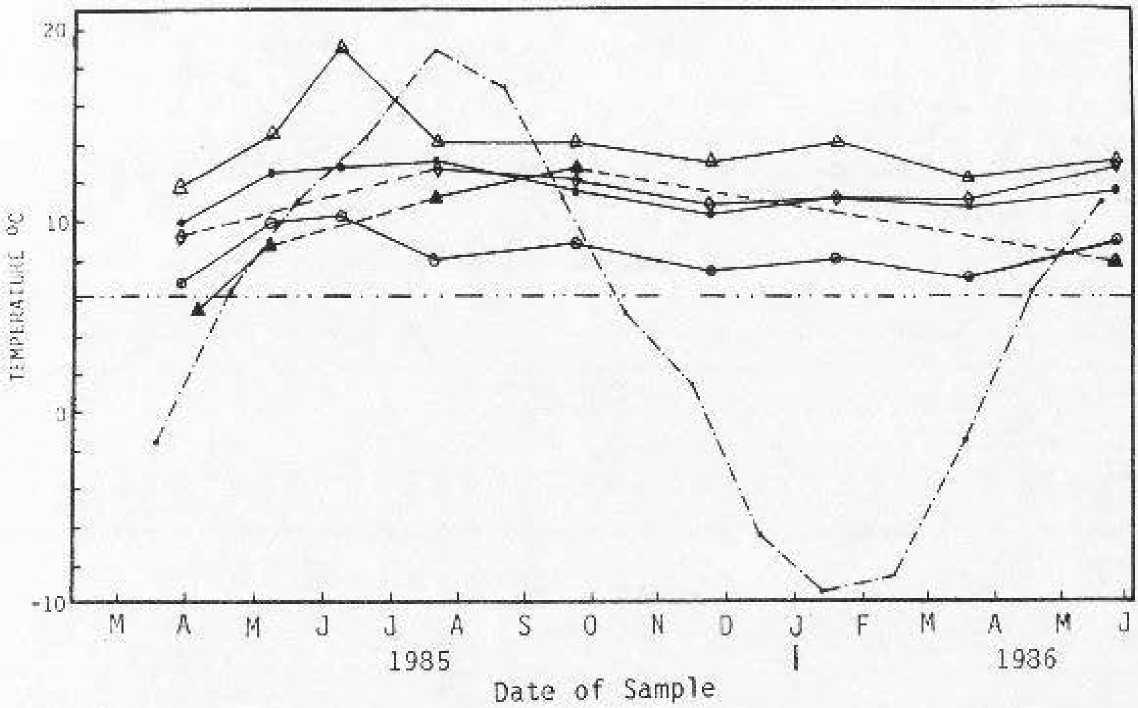
Hem (1970, p. 93) stated that the pH of most groundwater in the U.S. ranges from approximately 6.0 to 8.5, but thermal springs may yield water with lower pH values. In the study area, groundwater discharging from carbonate rocks and well-water had pH values ranging from 6.5 to 8.4. Spring-water from quartzites had pH values of 5.7 to 6.5 (appendix A).

Temperature

Temperatures of groundwater from carbonate springs in the study area range from 6.6° to 18.9° C, well-water ranges from 6.4° to 16.3° C, and groundwater from the quartzite spring ranges from 5.6° to 12.7° C (appendix A). Figure 4 shows that temperatures of Mantua Valley springs are warmer than the mean annual air temperature and do not seem to respond to average monthly temperature fluctuations.

Turbidity

Suspended and colloidal material in water, e.g., clay, silt, organic matter, and microscopic organisms, may cause water to be turbid (Todd, 1980, p. 290). Spring-water which deposited a yellowish residue on clean filters after two liters of water had been pumped through the filter, and/or spring-water which had a milky-white appearance, was considered to be turbid. Groundwater samples which showed signs of turbidity are summarized in table 4.



—·—·—Average Monthly Air Temperature; data are averages of daily temperature readings at the Hardware Ranch weather station [lat. 41°36' N., long. 111°34' W., elevation 1695 m (5560 ft)], and the Huntsville Monastery weather station [lat. 41°14' N., long. 111°43' W., elevation 1567 m (5140 ft)]; National Oceanic and Atmospheric Administration (1976-85)

—··—·· Mean Annual Air Temperature; data are averages of the average monthly air temperatures

----Missing data

Groundwater Temperature Readings:

- • Peter Jensen Spring
- ▵ West Hallings Spring
- ◊ Mud Spring
- ▲ Mantua Spring
- ⊙ Maple Spring

Figure 4.--Mantua Valley groundwater and average monthly air temperatures as functions of time.

Table 4.--Turbidity of groundwater samples
(0, not turbid; T, turbid; n.d., no data)

Station	May	Jun	Jul	Sep	Nov	Jan	Mar	May
Peter Jensen	0	0	0	0	0	0	0	0
West Hallings	0	0	0	0	0	0	0	0
Mud	n.d.	n.d.	0	0	0	0	0	0
Mantua	T	n.d.	0	T	n.d.	n.d.	n.d.	0
Maple	T	T	T	0	0	0	T	T
Well	T	T	0	0	0	0	0	0

The residue on the filters was examined by X-ray diffraction and with a binocular microscope for identification purposes. X-ray diffraction of a blank filter indicated that amorphous material was present. X-ray diffraction of the filtrate from both Maple and Mantua Springs revealed the presence of illite and/or montmorillonite as well as amorphous material. Maple Spring filter residue was transferred to a slide and examined under a 1000X binocular microscope. Unidentifiable angular particles and clumps of possible organic material were present (Larry Crist, Bio/West, Inc., oral commun., 1986). The owner of Mantua Fish Hatchery, which manages Maple Spring, reported that filters from turbid water discharging at low and high rates were examined under a microscope. The turbidity was the result of minute plant particles, which increased in concentration with increasing flow rates (Ron Roubidoux, Mantua Fish Hatchery, oral commun., 1985).

Primary Porosity

Results of porosity determinations for non-carbonate lithologies using the HCl acid test are presented in table 5.

Table 5.--Results of primary porosity estimates for non-carbonate rock samples

Formation	Description of Sample	Porosity (%)
Precambrian Caddy Canyon Quartzite	orange-tan, vitreous quartzite	<1
Precambrian Mutual Formation	purple, pebbly quartzite	<1
Cambrian Geertson Canyon Quartzite	tan quartzite	<1
Ordovician Swan Peak Formation	light-gray quartzite	<1
Mississippian Humbug Formation	tan, medium-grained, quartzitic sandstone	>12

All quartzite units analyzed for porosity were from highly fractured outcrops, but individual samples were clean and unweathered. The sample of the quartzitic sandstone was found as float where the formation was shown to crop out on the geologic map. Manger (1963) reported that the average porosity (determined with the bulk density-grain density method) of the Ordovician Swan Peak Formation from the Afton, Wyoming quadrangle is 2%.

Primary porosity of all the Paleozoic carbonate units which crop out in the study area is low (Peter T. Kolesar, Utah State University, oral commun., 1985). Point-counts of thin-sections of the Cambrian Langston and Ute Formations indicate that the primary porosities of those units are less than 1% (Rogers, 1987). Thin-section analysis of the Ordovician Garden City Formation indicates that the primary porosity is less than 1% (Susan K. Morgan, oral commun., 1986).

Fracture Analysis

Joints and bedding planes. Rose diagrams, representing joint and bedding plane strikes measured from outcrops in the study area, are presented on plate 2. Strikes of dominant joints in the northern portion of the study area generally range from N. 20° E. to N. 60° E., while joint orientations in the southern part are much more irregular. Nearly all outcrops examined have at least four or five joint sets with dips which intersect the dip of bedding at various angles. Of the 259 joint dips measured, 93% have dips equal to or greater than 45°.

Strikes of bedding planes are also more uniformly oriented in the northern part of the study area and are more irregular in the southern portion. Specific strikes and dips of bedding planes have been discussed in the Structure section.

Faults and lineaments. Lineaments, natural linear features, were identified from topographic maps and air-photos of the area and are presented on plate 2. Identification was subjective and involved the alignment of linear hillslopes, valley edges, saddles in ridges, and stream segments.

In the field, locations of mapped faults were examined for evidence of increased fracture spacing on either side of the fault. One of the examined faults, located in the saddle between Round and Gold Mine Hills, has an inferred orientation of N. 60° E. Two sets of joints (N. 23° E. and N. 85° W.) in Ordovician Swan Peak Formation outcrops south of the fault were analyzed (table 6). Distances between joints were measured perpendicularly to joint planes.

Table 6.--Fracture spacing (cm) south of fault between Round and Gold Mine Hills

Joint Orientation	Distance from Fault (m)			
	46	20	13	10
N. 23° E.	33	43	50	fragments, approx. 15 cm diam.
N. 85° W.	57	30	22	

Secondary porosity. Although secondary porosity could not be evaluated over the entire study area, several outcrops were analyzed to estimate the porosity that has developed due to fracturing and faulting. Secondary porosities of eleven formations in the study area (calculated by the method described in Data Collection) range from 1% to 23% (table 7) and are mainly the result of joint openings. Fractures in quartzites have straight walls with no evidence of mineralization or solution, whereas the majority of fractures in carbonate units show evidence of having been widened by solution. Calcite-filled veins (seam to 10 cm wide) were observed in several carbonate outcrops. Joint openings in the Langston, Blacksmith, and Swan Peak Formations are generally less than one cm wide and persist no deeper than 10 cm into the outcrop. Several joints in the Geertson Canyon, Bloomington, St. Charles, Garden City, Fish Haven, Laketown, and Brazer Formations are 15 to 61 cm (0.5-2.0 ft) wide at the surface of the outcrop. These wide joints often narrow 30% to 50% within one meter of the surface of the outcrop; occasionally apertures were found to exist at a depth of two meters into the

Table 7.--Estimates of secondary porosities in the Mantua Valley study area

Formation	Orientation of Joint Set	Estimated Porosity (%)[a]	Remarks
Cambrian Geertson Canyon Quartzite	N. 75° E., 85° SE. N. 60° W., 50° SW.	6 8	well-fractured roadcut
Cambrian Langston Dolostone	N. 25° W., 55° NE.	3	avg. width 0.25 cm
Cambrian Blacksmith Dolostone	N. 80° E., 74° NW. N. 70° W., 73° NE.	10 10	closely spaced joints; apertures diminish quickly with depth
Cambrian Bloomington Formation	N. 60° W., 85° NE.	23	avg. width 7.5 cm; some joints open up to 46 cm at surface
Cambrian St. Charles Dolostone	N. 20° W., 48° SW.	11	avg. width 5 cm; some joints open up to 46 cm at surface
Ordovician Garden City Formation	N. 85° E., 70° SE.	17	avg. width 5 cm; some joints open up to 15 cm at surface
Ordovician Swan Peak Formation	N. 23° E., 70° NW. N. 85° W., 75° SW.	2 1-3	avg. width 0.75 cm; near a fault
Silurian Laketown Dolostone	N. 10° E., 85° NW. N. 10° W., 85° SW.	12 14	avg. width 7.5 cm; some joints open up to 46 cm at surface
Mississippian Lodgepole Limestone	N. 40° W., 74° NW.	8	avg. width 2.5 cm
Mississippian Humbug and Deseret Formations, undivided	N. 20° W., 85° SW. N. 10° E., 65° NW.	11 14	avg. width 2.5 cm; avg. width 4.6 cm
Mississippian Brazer Formation	N. 40° W., 52° SW. N. 45° E., 74° SE.	11 10	avg. width 2 cm; some joints open up to 30 cm at surface

[a]see appendix B for data used to make porosity estimates

outcrop. Some outcrops of the Blacksmith, Bloomington, St. Charles, Garden City, Laketown, and Brazer Formations have bedding planes which have been widened by solution. One Laketown outcrop has an estimated porosity of 7% due to separation between bedding planes.

Laboratory Analysis

A check on the precision of the laboratory analyses was performed by analyzing the same sample five or six times and computing the mean and standard deviation of the duplicate analyses. Results of the reproducibility tests are shown in table 8.

Table 8.--Reproducibility tests of laboratory analyses

Ion/Species	Number of Analyses	Mean (mg/l)	Standard Deviation (mg/l)
Sodium	6	8.8	0.1
Potassium	6	1.1	0.0
Calcium	6	46.6	0.4
Magnesium	6	17.9	0.2
Chloride	5	8	0.3
Sulfate	4	6	0.0
Silica	5	12	0.0
TDS	5	239	5.4

The accuracy of the chemical analyses was checked by calculating cation-anion balances. When concentrations of ions are converted to milliequivalents per liter (meq/l), the sum of the cations should approximately equal the sum of the anions. Hem (1970, p. 234) stated that differences between cations and anions larger than 2% are sometimes unavoidable when the total of cations and anions is less than 5.00 meq/l. Kroneman, at the Earth Science Lab in Salt Lake City, Utah, reported by de Vries (1982), said that a 10% difference between cations and anions is acceptable, 15% questionable, and greater than 15%, unacceptable. Values obtained from the WATEQF program (Plummer and others, 1976) for HCO_3^- and CO_3^{-2}, and values obtained for Ca^{+2}, Mg^{+2}, Na^+, K^+, SO_4^{-2}, and Cl^- in the lab were used to compute the cation-anion balances. Results of the accuracy checks are presented in table 9.

Chemical Character of Groundwater

Concentrations of Na^+, K^+, Ca^{+2}, Mg^{+2}, Cl^-, SO_4^{-2}, HCO_3^-, CO_3^{-2}, SiO_2, and TDS were determined for the samples on a monthly to bimonthly basis, and the results are presented in appendix A. Chemical constituents in Mantua Valley springs are derived principally from solution of carbonate minerals in limestones and dolostones and possibly from solution of siliceous microfossils in those units. Mean ion concentrations for each spring are presented in table 10.

Groundwater emerging as springs from carbonate rocks has high concentrations of Ca^{+2} and Mg^{+2} and therefore high hardness. Spring-water emanating from quartzites has very low concentrations of all ions and exceptionally low TDS. Average concentrations of ions of each spring, (converted from milliequivalents to percent) plotted on a trilinear diagram (Piper, 1944), graphically demonstrate the chemical nature of Mantua Valley groundwater (fig. 5). Figure 5 shows that all the groundwater sampled is of the calcium-magnesium-bicarbonate type.

Table 9.--Accuracy tests of laboratory analyses

Sample	Cations (meq/l)	Anions (meq/l)	Difference (%)
Peter Jensen			
Jul	4.52	4.10	5
Sep	4.57	4.32	3
Nov	4.50	4.75	3
Jan	4.30	5.01	8
Mar	4.49	4.47	0.2
May	5.04	4.99	0.5
West Hallings			
Jul	3.99	4.40	4
Sep	4.30	3.97	4
Nov	4.19	4.54	4
Jan	3.99	4.49	6
Mar	4.20	4.38	2
May	4.79	4.60	2
Mud			
Jul	4.33	4.38	1
Sep	4.32	4.29	0.3
Nov	4.36	4.05	4
Jan	4.29	4.43	2
Mar	4.29	4.60	3
May	4.96	4.39	6
Mantua			
Jul	0.69	0.87	12
Sep	0.71	0.89	12
Nov	1.25	1.56	11
Jan	1.39	1.72	11
Mar	2.78	3.19	7
May	0.66	0.68	1
Maple			
Jul	3.40	4.14	10
Sep	3.68	3.85	2
Nov	3.60	3.72	2
Jan	3.43	3.96	7
Mar	3.14	3.54	6
May	3.77	3.64	2
Well			
Jul	5.31	6.06	7
Sep	5.17	5.35	2
Nov	5.21	5.32	1
Jan	5.09	6.01	8
Mar	5.26	5.07	2
May	6.02	5.56	4

Table 10.--Mean ion concentrations of Mantua Valley groundwater (mg/l)

Station	Ca^{+2}	Mg^{+2}	Na^{+}	K^{+}	HCO_3^{-}	SO_4^{-2}	Cl^{-}	SiO_2	TDS[a]
Peter Jensen	53.1	19.0	7.3	0.8	260	9	8	8.8	256
West Hallings	44.3	19.8	8.5	1.1	245	8	11	11	247
Mud	49.3	20.7	6.7	1.0	246	9	8	9.4	251
Mantua	7.0	2.0	4.0	0.8	39	6	3	7.4	65
Maple	39.0	14.4	6.6	0.9	217	7	6	8.4	193
Well	57.3	22.7	14.1	1.1	315	8	12	14	309

[a]as residue on evaporation

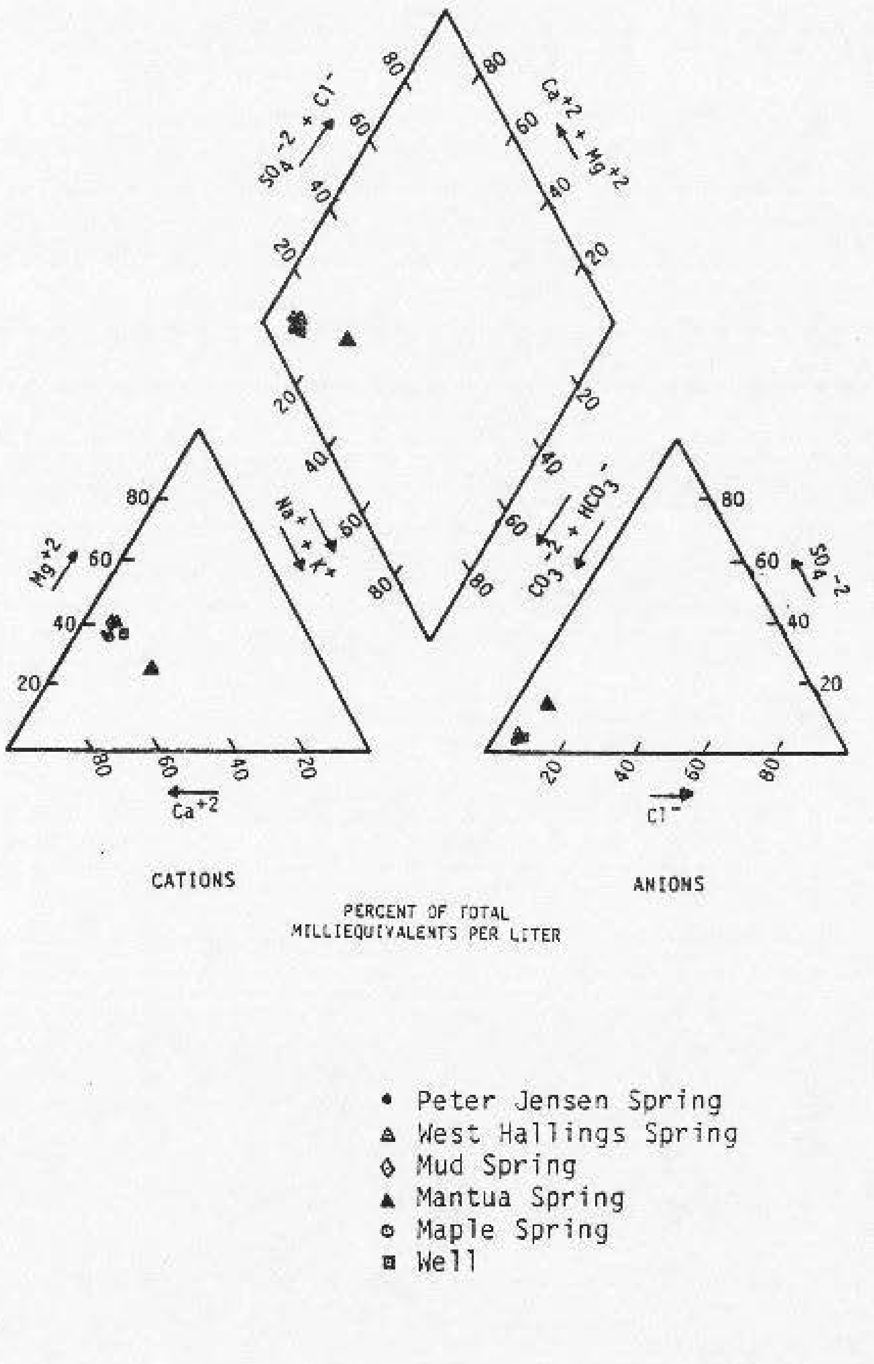


Figure 5.--Trilinear diagram (Piper, 1944) of average ion concentrations of Mantua Valley groundwater.

Cations. Dominant cations in groundwater issuing from carbonate units and the well are Ca^{+2} and Mg^{+2}, while Na^{+} and K^{+} are much less abundant. Groundwater flowing from quartzites has very low concentrations of cations, with Ca^{+2} the most abundant, followed by Na^{+}, Mg^{+2}, and K^{+}. Very low concentrations of dissolved iron are also present in Mantua Valley groundwater (Bjorklund and McGreevy, 1973).

Hem (1970, p. 131) stated that Ca^{+2} is the principal cation in most natural fresh water, and Mantua Valley groundwater is no exception. Groundwater issuing from carbonate rocks has from 31.6 to 59.4 mg/l Ca^{+2}, and groundwater from the well has 53.3 to 65.0 mg/l Ca^{+2}. Groundwater flowing through quartzites has Ca^{+2} concentrations of 6.3 to 7.7 mg/l. Ca^{+2} concentrations of each spring, as functions of time, are shown on figure 6.

Concentrations of Mg^{+2} in most natural fresh water are much lower than concentrations of Ca^{+2} (Hem, 1970, p. 144). Figure 6 shows that concentrations of Mg^{+2} in Mantua Valley groundwater range from 13.4 to 22.5 mg/l in carbonate springs, from 19.3 to 25.7 mg/l in the well, and from 1.8 to 2.1 mg/l in the quartzite spring.

Na^{+} and K^{+} are both members of the alkali-metal group. Hem (1970, p. 150) stated that concentrations of Na^{+} are higher than concentrations of K^{+} in most natural water, and again Mantua Valley groundwater displays typical concentrations of cations. Figure 6 shows that groundwater emanating from carbonate rocks has 5.6 to 10.0 mg/l Na^{+} and 0.7 to 1.8 mg/l K^{+}. Well-water ranges from 13.3 to 15.6 mg/l Na^{+} and from 0.7 to 2.3 mg/l K^{+}. Groundwater discharging from quartzites has 3.6 to 4.5 mg/l Na^{+} and 0.7 to 1.5 mg/l K^{+}.

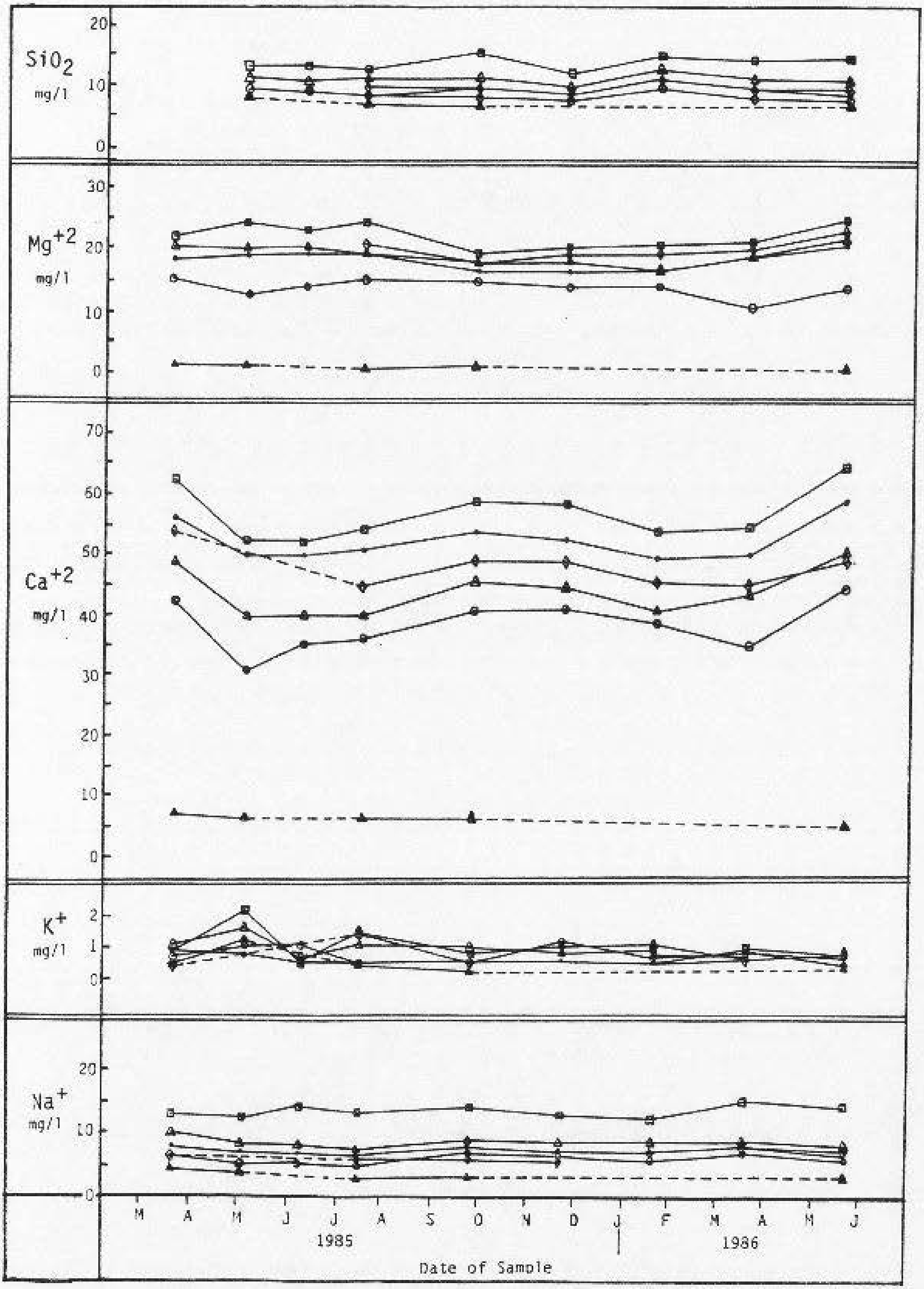


Figure 6.--Cations and silica concentrations as functions of time. •, Peter Jensen Spring; △, West Hallings Spring; ◇, Mud Spring; ▲, Mantua Spring; ⊙, Maple Spring; ⊡, Well; ---, missing data.

Bjorklund and McGreevy (1973, p. 20-22) have reported chemical analyses indicating that dissolved iron concentrations are very low in Mantua Valley groundwater. Mantua Valley spring- and well-water, sampled in the mid- to late-1960's, had 0.00 to 0.01 mg/l dissolved iron. Water sampled from Mantua Spring in 1941 had 0.05 mg/l dissolved iron.

Anions. The dominant anion of all Mantua Valley groundwater is HCO_3^-, while Cl^- and SO_4^{-2} are much less abundant. Cl^- and SO_4^{-2} are present in approximately equal amounts in the carbonate springs, while SO_4^{-2} is slightly more abundant than Cl^- in the quartzite spring. Very low concentrations of NO_3^-, B, and F^- (Bjorklund and McGreevy, 1973), and CO_3^{-2} are also present in Mantua Valley groundwater.

Alkalinity refers to the ability of waters to neutralize acids and is dependent upon dissolved species of $CaCO_3$ and pH. Alkalinity in most natural water is primarily represented by dissolved HCO_3^- and CO_3^{-2} (Hem, 1970, p. 157). Alkalinity in Mantua Valley groundwater is almost entirely produced by HCO_3^- ions, since CO_3^{-2} ions are present in concentrations of 1 mg/l or less. Figure 7 shows that HCO_3^- concentrations in groundwater flowing from carbonate units range from 200 to 284 mg/l. Well-water has 284 to 342 mg/l HCO_3^-, and groundwater from quartzites has 33 to 42 mg/l HCO_3^- (fig. 7).

Sulfate (SO_4^{-2}) is fully oxidized sulfur (S^{+6}) complexed with oxygen (Hem, 1970, p. 161). Figure 7 shows that concentrations of SO_4^{-2} in Mantua Valley groundwater are relatively constant. Water from carbonate springs has 5 to 11 mg/l SO_4^{-2}, and well-water has 7 to 10 mg/l SO_4^{-2}. Groundwater emanating from quartzites has SO_4^{-2} concentrations of 4 to 6 mg/l (fig. 7).

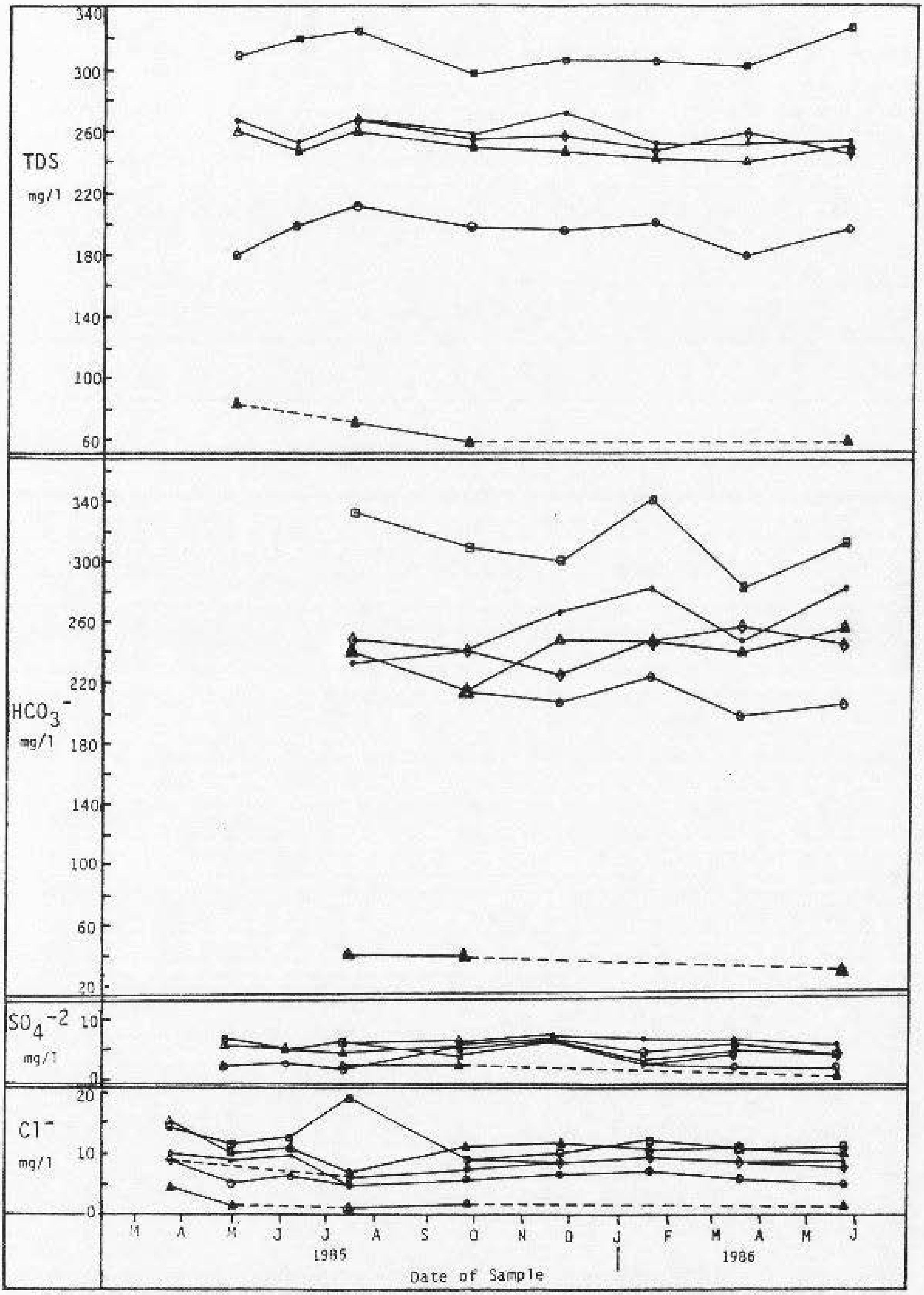


Figure 7.--Anions and TDS concentrations as functions of time. •, Peter Jensen Spring; △, West Hallings Spring; ◇, Mud Spring; ▲, Mantua Spring; ⊙, Maple Spring; ⊡, Well; ---, missing data.

Chloride (Cl^-) is present, usually in low concentrations, in all natural waters (Hem, 1970, p. 172), and Mantua Valley groundwater is no exception. Cl^- concentrations in groundwater discharging from carbonate units range from 5 to 15 mg/l, and well-water has from 8 to 18 mg/l Cl^- (fig. 7). Groundwater from quartzites has 2 to 4 mg/l Cl^- (fig. 7).

Bjorklund and McGreevy (1973, p. 20-22) have reported concentrations of NO_3^-, B, and F^- in carbonate spring- and well-water sampled in the mid- to late-1960's. NO_3^- concentrations were 4.5 mg/l or less, B concentrations were 0.17 mg/l or less, and F^- concentrations were 0.2 mg/l or less. The U.S. Environmental Protection Agency (1986) has set the F^- secondary maximum contaminant level for public water systems at 2.0 mg/l. Because Mantua Valley groundwater has low concentrations of F^-, the Brigham City Water Department fluoridates its water for municipal use (Joseph P. Marshall, Brigham City Corp., oral commun., 1985). The spring emanating from quartzite units, sampled in 1941, had a concentration of 0.2 mg/l F^-.

Silica. Silicon in natural waters is actually in a hydrated form and is represented as H_4SiO_4 or $Si(OH)_4$, however the term "silica" refers to the oxide SiO_2 and represents silicon in natural water (Hem, 1970, p. 103). Figure 6 shows that silica ranges from 7.6 to 13 mg/l in groundwater from carbonate rocks, 12 to 15 mg/l in well-water, and 7.1 to 7.8 mg/l in groundwater flowing through quartzites.

Total dissolved solids. Total dissolved solids concentrations are a practical way of comparing water types (Hem, 1970, p. 219). TDS of groundwater flowing through carbonate rocks ranges from 176 to 268 mg/l, well-water has from 294 to 324 mg/l, and groundwater from quartzites has from 56 to 80 mg/l TDS (fig. 7). Water with less than 1000 mg/l TDS is classified as fresh water (Robinove and others, 1958).

Saturation Indices

The saturation index (SI) of a given mineral is equal to the log of the quotient of the ion activity product and the solubility product, i.e., SI = log (IAP/Ksp). A saturation index does not indicate that the mineral is actually present in the groundwater, but it is indicative of the state of equilibrium, if, in fact, the mineral were present. Saturation indices are presented in logarithmic form and have one of three interpretations: 1) SI values equal to 0.00 ±0.1 suggest equilibrium, i.e., water that is saturated with respect to the specified mineral; 2) negative SI values indicate undersaturated water, i.e., water that will dissolve the specified mineral if encountered; and 3) positive SI values indicate supersaturated water, i.e., water that will precipitate the specified mineral under favorable pressure and temperature conditions.

Mantua Valley groundwater is substantially undersaturated in some samples with respect to calcite and dolomite and is supersaturated in all samples with respect to quartz. Saturation indices of calcite, dolomite, and quartz for each sample are presented in appendix C. Averages of six saturation indices for each spring, for calcite, dolomite and quartz, are presented in table 11. Based on an uncertainty of ±0.1 (Langmuir, 1971), one sample from Mud Spring reached saturation with respect to dolomite, and one sample from Mud Spring and three from the well were supersaturated with respect to dolomite (appendix C). Figure 8, which shows saturation indices with respect to calcite as functions of time, indicates that Maple Spring never reached saturation, and Peter

Table 11.--Average saturation indices of Mantua Valley groundwater

Station	Calcite	Dolomite	Quartz
Peter Jensen	-0.372	-1.050	0.424
West Hallings	-0.286	-0.752	0.454
Mud	-0.230	-0.683	0.410
Mantua	-2.854	-6.132	0.320
Maple	-0.453	-1.271	0.425
Well	-0.059	-0.413	0.600

Jensen Spring reached saturation once. West Hallings and Mud Springs reached saturation with respect to calcite twice, Mud Spring was supersaturated once, and the well was supersaturated three times (fig. 8).

Tritium Analyses

Tritium is the radioactive isotope of hydrogen with a mass of three (^3H) and with a half-life of 12.43 years, i.e., the decay constant, λ = 5.576%/year (H. Gote Ostlund, Tritium Laboratory, University of Miami, written commun., 1986). Tritium is utilized in scientific studies in two ways: 1) "environmental tritium" normally occurs in waters, and because it decays radioactively, it can be used to date young (<50 years old) groundwater; and 2) tritium, when artificially introduced into the hydrologic cycle, may be used as a tracer. The tritium referred to in this report is "environmental tritium".

Activities (concentrations) of tritium are reported as tritium units (TU), where one TU is equal to one tritium atom per 10^{18} hydrogen atoms. Tritium is produced by cosmic radiation and is present in atmospheric

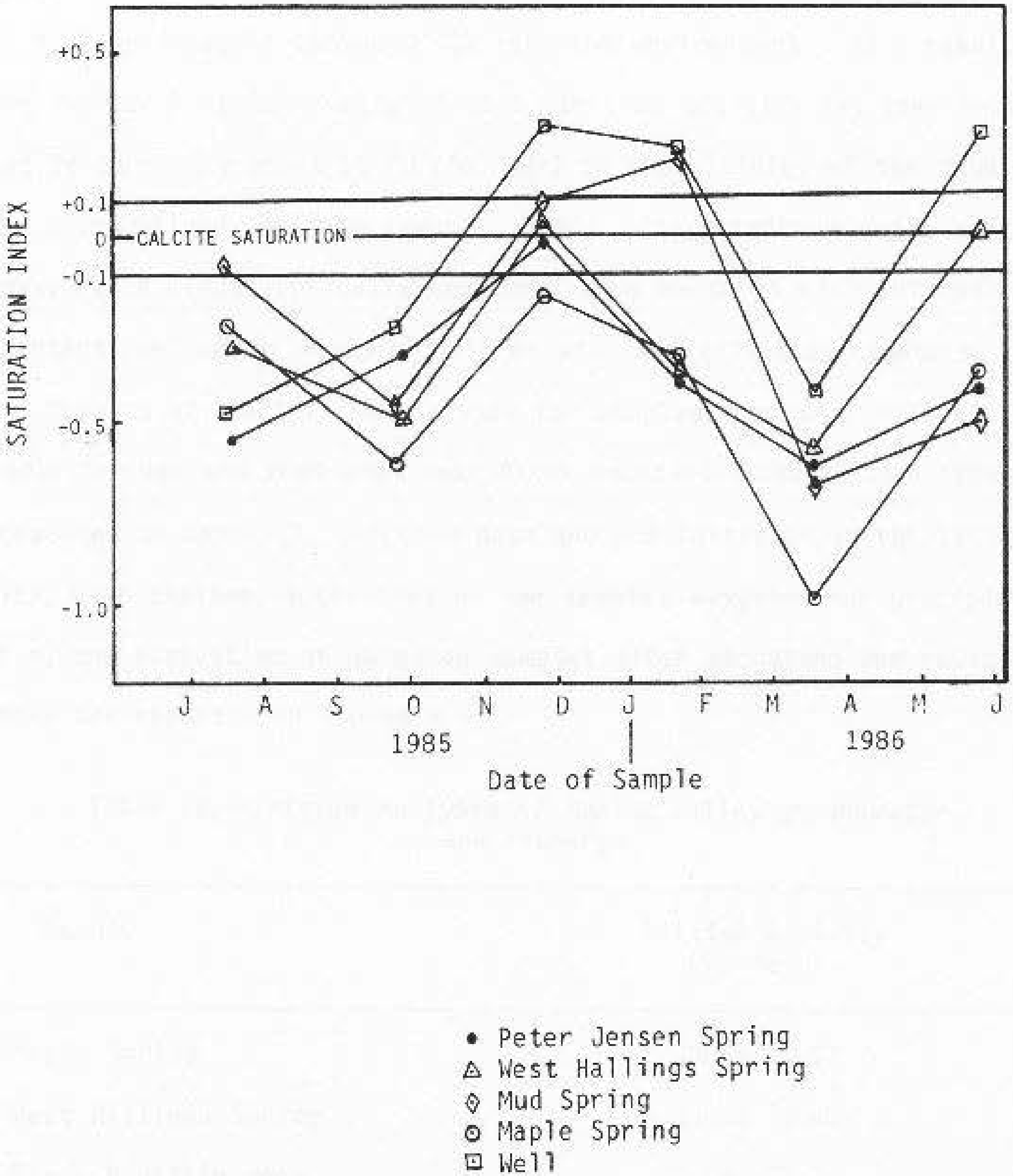


Figure 8.--Saturation indices with respect to calcite as functions of time.

water in concentrations of approximately 10 TU. Atmospheric nuclear tests, starting in 1954 and peaking in 1963, introduced large quantities of tritium (several thousand TU) into the environment. As a result of the Nuclear Test Ban Treaty of 1963, tritium activity has been declining, and is currently about 10 TU (in rain) in the vicinity of the study area (H. Gote Ostlund, written commun., 1986). Since tritium emits soft beta rays, it is electrolytically enriched, then measured with internal gas counters (Geiger or proportional) or with scintillation counters.

Results of the tritium analyses for samples from West Hallings and Maple Springs and from snow near Black Mountain taken 29 June 1986 are presented in table 12. Tritium data and precipitation at the Salt Lake City, Utah station, activities of the samples weighted for precipitation, and activities of weighted samples after adjusting for radioactive decay are reported in appendix D.

Table 12.--Tritium analyses of Mantua Valley groundwater and recharge

Sample	Tritium Activity (TU ±eTU)
Maple Spring	26.05 ±0.61
West Hallings Spring	16.02 ±0.46
Black Mountain snow	8.86 ±0.24

INTERPRETATION

Interpretations of the hydrogeology of Mantua Valley and the surrounding area were based on field and laboratory data. Field data include fracture orientations, lineaments, stratigraphic, structural, and topographic relationships, joint spacing and apertures, groundwater temperature data, and the relative amounts of solution observed along bedding planes and joints of carbonate units. Laboratory data include Ca^{+2}/Mg^{+2} ratios, TDS concentrations, saturation indices, CO_2 partial pressures, and tritium analyses.

Hydrostratigraphic Units

Geologic formations in the study area were divided into aquifers and aquitards based on secondary porosity estimates made from the spacing and openings of joints and bedding planes, field observations of the relative amounts of solution along bedding planes and joints, the magnitudes and numbers of springs discharging from the units, and the lithologies of the units. Total thickness of aquitards in the study area ranges from 2815 to 3670 m, while total aquifer thickness ranges from 3625 to 4253 m.

The Precambrian units, the Cambrian Geertson Canyon Quartzite, and the Ordovician Swan Peak Formation are thought to be aquitards. Although these quartzites are all highly fractured, the fractures are generally narrow even at the surface. The fractures have not been widened by solution and probably do not extend to great depths. However, lineaments in these units may indicate localized zones of higher permeability at depth. The Precambrian and Cambrian quartzites constitute the Precambrian and Cambrian quartzite aquitard of plate 1, while the Ordovician Swan Peak Formation composes the Lower Paleozoic quartzite aquitard of plate 1.

The Cambrian Langston, Ute, and Blacksmith Formations are classified as aquitards based on the following field evidence: 1) all three formations have generally very narrow fractures in outcrops; 2) little evidence of solution was observed along joints or bedding planes in outcrops of the three formations; 3) the Langston and Ute Formations have shale and siltstone interbeds; and 4) no known springs discharge from the three formations in the study area. The Langston, Ute, and Blacksmith Formations compose the Lower Paleozoic carbonate aquitard of plate 1.

The Cambrian Bloomington Formation must be an aquifer since two large springs (West Hallings and Maple Springs) discharge from the unit in the study area. However, the Hodges Shale Member of the Bloomington Formation may act as an aquitard because of its fine-grained texture. Extensive solution of bedding planes and joints was observed in outcrops of the Bloomington Formation. The Bloomington Formation constitutes part of the Lower Paleozoic carbonate aquifer of plate 1.

The Cambrian Nounan and St. Charles Dolostones and the Ordovician Garden City Formation are aquifers. A large spring (Mud Spring) discharges from the St. Charles Dolostone in the study area, and solution along bedding planes and joints was observed in outcrops of all three formations. Several small sinkholes were observed in alluvium overlying the Ordovician Garden City Formation. The Nounan, St. Charles, and Garden City Formations constitute part of the Lower Paleozoic carbonate aquifer of plate 1.

The Ordovician Fish Haven and Silurian Laketown Dolostones are probably aquifers. They compose the Middle Paleozoic carbonate aquifer of plate 1. Very large joints and bedding planes which have been enlarged by solution were observed in outcrops of these two formations.

The Devonian units, the Water Canyon Formation and the Hyrum Dolostone, are probably aquifers, although some silty beds may act as aquitards. The Devonian units are not widely exposed in the study area, but the outcrops which were observed showed evidence of solution along bedding planes and joints. The Water Canyon Formation and the Hyrum Dolostone compose part of the Upper Paleozoic carbonate aquifer of plate 1.

The Mississippian units, the Lodgepole and Deseret Limestones and the Humbug and Brazer Formations, are aquifers. A number of springs along the east side of Mantua Valley discharge from, and probably flow through, several of the Mississippian units. A few small sinkholes in alluvium overlying the Brazer Formation were observed in the study area. Joints and bedding planes which have been widened by solution were observed in all four of the Mississippian units. The Mississippian units constitute the majority of the Upper Paleozoic carbonate aquifer of plate 1.

The Tertiary Wasatch and Evanston Formations may be aquifers in places but probably contain beds which may behave as aquitards because of cementation. The Tertiary units compose the Tertiary aquifer and aquitard of plate 1. All of the Quaternary units in the study area are probably aquifers, because of the generally coarse textures, although permeabilities may decrease locally because of clay beds or other impermeable lenses. The Quaternary units constitute the Quaternary aquifer of plate 1.

Origin of Topographic Features

The valley in which Dry Lake is situated is closed to surface-water drainage. In addition to silt and clay deposits, abundant coarse-grained gravels, transported from the Wasatch Mountains, contribute to the Dry Lake Valley fill (Williams and others, 1970). Depth to bedrock beneath Dry Lake, as well as most of the valley, is less than 12 m (40 ft; Williams and others, 1970). The elongate valley measures roughly 914 m (3000 ft) by 3048 m (10,000 ft).

A possible explanation for the closed nature of Dry Lake Valley is that it is a down-dropped fault block, closed to surface-water drainage by alluvial fan deposits. Dry Lake Valley trends roughly north-south, as do larger fault-block valleys in the region (Cache and Bear River Valleys). Air photos suggest that the northern edge of Dry Lake Valley is presently closed by a large alluvial fan, the material of which was/is derived from Snow Canyon. Williams and others (1970) indicate that coarse alluvial deposits in the fan are as much as 36.5 m (120 ft) thick. The lowest point along the northern Dry Lake Valley boundary, which according to Williams and others (1970) corresponds to an alluvial thickness of 36.5 m (120 ft), is approximately 1749.6 m (5740 ft). If bedrock is at approximately 1713 m (5620 ft), i.e., 36.5 m (120 ft) below the present surface, surface-water may originally have drained from Dry Lake Valley [bedrock at approximately 1716 m (5630 ft)] northward. Because the bedrock gradient is so low, and perhaps because the valley was a groundwater discharge area, water may have ponded at times in the valley. Water percolating along joints and bedding planes of the carbonate rocks which crop out in the valley may have widened these features by solution. An exceptionally large debris flow event or events could have blocked the Dry Lake Valley outlet, closing the valley to surface-water drainage. Normally such a blockage would be removed by overtopping, but previously-formed solution conduits located on the east side of Dry Lake may have provided a subsurface outlet for excess water.

For the past several years, Dry Lake Valley has perennially contained ponded water. The lake level decreases by evaporation, drainage through small sinkholes on the east side of the lake, and possibly by migration of water through the coarse alluvial fan deposits at the north end of the valley. An increase in precipitation over the area in recent years, as well as subterranean drainage passages choked with clay, silt and debris, may account for the recent perennial nature of Dry Lake.

Mantua Valley is another elongate valley, roughly 2286 m (7500 ft) wide and 3658 m (12,000 ft) long. Drill logs of water wells in Mantua Valley (Bjorklund and McGreevy, 1973) indicate that the valley contains at least 154 m (505 ft) of surficial fill. Because bedrock is exposed in Box Elder Creek Canyon at an elevation of 1560 m (5120 ft), and because bedrock beneath Mantua Valley is at an elevation less than 1425 m (4675 ft), the valley is a closed bedrock depression and may have, at one time, been closed to surface-water drainage. Other drill logs of water wells in the valley (Bjorklund and McGreevy, 1973; Well Driller Reports, on file in the Office of State Engineer), demonstrate the relatively rhythmic layering of the valley fill. The sediments filling the valley are predominantly clays, gravels, and boulders. The upper portion of the valley fill was presumably deposited into Lake Bonneville during its highest stand. Underlying the lake deposits is valley fill, which was probably deposited by Box Elder Creek and other streams at an earlier time. The large closed depression may have been the result of karst activity, which may have caused a very large sinkhole to develop and subsequently collapse. However, it is improbable that a sinkhole would have developed across the ten stratigraphic units which are exposed to the north and south of the valley. Also, if a sinkhole of such magnitude had developed, Mantua Valley and vicinity would probably be underlain by an extensive network of caverns, some of which would most likely be evident in the area today. One opening, which appears to be relatively shallow, is visible in cliffs of the Ordovician Garden City Formation above Devils Hole Canyon. The opening could be the result of past groundwater flow systems, or it could have been created by the erosive action of Box Elder Creek as it downcut through bedrock at that elevation. No other known caves exist in the study area. The best explanation for Mantua Valley is a tilted block. Faulting is the probable origin of numerous other north-south-trending, linear valleys in the region.

Clay Valley is an elongate valley, approximately 1524 m (5000 ft) by 2743 m (9000 ft), which trends northeast-southwest. A bordering fault is evident along one side of Clay Valley (Martin L. Sorensen, U.S. Geological Survey, written commun., 1986). Air-photo analysis suggests that Clay Valley may have once been a closed drainage basin, since it has a relatively flat floor and high bordering ridges. However, a headward-eroding stream from the west appears to have captured the drainage through a steep narrow divide and incorporated it into the Mantua Valley drainage system. The fact that relatively little incision of the Clay Valley fill is observed, and that the capturing stream lies in such a narrow, steep, and rugged canyon, suggests that the capture may have occurred in the relatively recent past (i.e., the Holocene).

Jepsen Valley, located slightly northwest of Mantua Valley, seems to have had a recent geologic history similar to Clay Valley. Jepsen Valley appears to have also been a closed depression, only recently captured by a small headward-eroding stream.

Sink Hole Valley, an oval-shaped valley roughly 1524 m (5000 ft) by 3048 m (10,000 ft), is closed to surface-water drainage. The valley may have been closed as a result of normal faulting, since a normal fault borders the east side of the valley (Crittenden and Sorensen, 1985). At least three sinkholes, approximately 4.5 m (15 ft) in diameter and 3 m (10 ft) deep, which are drained through small holes in the surficial cover, are present in the extreme southwest corner of the valley. A mapped fault in the southwest corner of Sink Hole Valley (Crittenden and Sorensen, 1985) suggests that the sinkholes may be related to fault-fractured rock. A narrow, elongate projection of Mississippian bedrock extends approximately 1524 m (5000 ft) from the western border of Sink Hole Valley toward the center of the valley (pl. 1). The Mississippian rocks consist of limestone, dolostone, chert, quartzitic sandstone, and silicified fossils. The silica content of the Mississippian rocks probably makes the outcrops more resistant to erosion and solution. Air-photo analysis suggests that the valley will eventually be captured by headward-eroding drainages along the southwestern or the eastern boundary.

Devils Gate Valley, in the southernmost portion of the study area, is a large oval-shaped valley, measuring approximately 3658 m (12,000 ft) by 4572 m (15,000 ft). The surficial fill in Devils Gate Valley has been deeply incised by the headwaters of Box Elder Creek. By using a Brunton pocket transit as a level and traversing a bank of one of the deepest tributaries of Box Elder Creek, the surficial fill of Devils Gate Valley was found to be at least 30.5 m (100 ft) thick. These deep headwater drainages are separated by undissected areas, which appear to be remnants of an older surface. The level nature of this remnant surface may indicate that Devils Gate Valley was a closed depression before capture by Box Elder Creek. Air-photo analysis shows that Box Elder Creek is rapidly incising into bedrock in Devils Hole Canyon in sec. 3, T. 8 N., R. 1 W., and the upstream tributaries in Devils Gate Valley are responding by down-cutting. Devils Gate Valley may be underlain by a series of tilted blocks (Crittenden and Sorensen, 1985, cross-section B-B'). Davis (1985) suggested that the valley was formed in part by solution of the Garden City and underlying formations.

Present day sinkholes in and around the study area are relatively small, e.g., 4.5 m diameter by 3 m deep (15 ft by 10 ft). Karst processes in the region are probably not very active because of the relatively low precipitation and the lack of thick, humus-rich soils in the area. However, at least some solution of carbonate rocks should be occurring, because much of the groundwater in Mantua Valley is undersaturated with respect to calcite and dolomite. The study area may have experienced stronger karst activity in the Tertiary when temperatures were warmer, precipitation was higher, and general uplift of the region created steep hydraulic gradients. Wilson (1976) reported that primary karstification in the Bear River Range occurred prior to Bull Lake Glaciation, and possibly as much as 350,000 years before present. A large hole in bedrock on the east side of a U.S. Route 89-91 roadcut in the northern part of sec. 15, T. 9 N., R. 1 W. is filled with Wasatch Formation and appears to be a possible sinkhole remnant. However, careful field examination of the hole reveals that it is located along a northwest-trending linear feature, which exhibits a zone of anomalously small clasts at the ground surface, suggesting a possible fault-related origin. Locations of several small sinkholes/closed depressions which were located during air-photo analysis and field work are shown on plate 2. Larger sinkholes are known in the Bear River Range, at higher elevations than the study area, and may be a result of greater amounts of precipitation and steeper hydraulic gradients in the Bear River Range.

The present topography in the study area is probably chiefly the result of a complex structural history. Basin and Range extension, which continues to the present in the region, has created numerous down-dropped blocks of various sizes, which in turn have become the sites of valleys. Many of the down-dropped blocks in the study area may once have been closed depressions. Small-scale anomalous topographic features may have been formed by groundwater circulation and solution along zones of fault-fractured rock. The existence of a large closed depression, Sink Hole Valley, and the fact that extremely active solution of bedrock is currently not evident, suggests that karst activity is not responsible for much of the anomalous topography in the study area.

Porosity and Permeability

Primary porosities of most of the formations in the study area are extremely low. If primary porosity is essentially nonexistent, then permeability due to the interconnectedness of pore spaces is probably low as well. Even though the quartzitic sandstone unit of the Mississippian Humbug Formation has a relatively high primary porosity (table 5), this unit alone cannot account for high groundwater discharge into Mantua Valley, since it occupies a relatively small portion of the study area. Conditions other than primary porosity in rocks must be allowing groundwater flow within the study area.

Secondary porosity and permeability can develop in rocks through faulting, jointing, and solution of material. The region in which the study area is situated has experienced intense structural deformation from the Early Cretaceous to the present, including compression, uplift, and extension. Such structural deformation probably accounts for the highly faulted and jointed nature of the rocks in the study area. The Precambrian and Cambrian quartzites have been especially affected by deformation. Outcrops of these brittle formations are generally highly fractured rather than massive, cliff-forming walls. The carbonate units in the study area have been affected to varying degrees by deformation, but all are jointed. However, some formations display wider and deeper joints than others. Larger joints in some carbonate formations may be due to inherent weakness within the formation and/or to modification by solution. As discussed in Previous Investigations, a number of the Paleozoic carbonate units exposed in the study area are known to have experienced karstification. It is evident from table 7 that outcrops of some formations in parts of the study area have adequate secondary porosities, in the form of joints, for groundwater flow. Davis (1969) reported that in general, the porosity and permeability of originally dense rocks should be about ten times greater at a depth of 10 m than at a depth of 100 m. It has been shown by Gale (1982) that permeability of fractured rocks commonly decreases with depth, therefore it should be expected that the secondary porosity values reported in table 7 would also decrease with depth. However, bedding planes and/or joints may have been enhanced by solution, since the majority of Mantua Valley groundwater is undersaturated with respect to calcite and dolomite (appendix C). If some of the joints widened by solution intersect bedding planes or joints widened by solution at depth, it is reasonable to conclude that such secondary permeability is an important factor contributing to groundwater flow within the study area.

Whether or not faults contribute directly to groundwater flow is not obvious from field evidence. Three of the Mantua Valley springs are located near inferred faults (pl. 1). Field observations indicate that between Gold Mine and Round Hills, the fracture spacing of one joint set increases with proximity to an inferred fault (table 6). This increase in the density of fracturing implies an increase in the permeability at depth. Groundwater has the potential to flow along fault planes, especially if the faults have been widened by solution in carbonate units. However, some faults may be plugged by gouge or material precipitated by groundwater and therefore would no longer provide significant avenues for groundwater flow.

Depth of Groundwater Circulation

Monthly to bimonthly temperature readings of groundwater were used to approximate the depth of circulation of groundwater supplying the springs to further characterize the groundwater flow systems. The absence of large fluctuations in temperatures and the fact that the temperatures are warmer than the mean annual air temperature (fig. 4) suggest that the groundwater flows deeply enough to be warmed by the geothermal gradient. The average geothermal gradient along the Wasatch Front for carbonate rocks is 30° C/km, for thick quartzites, it is 18° C/km, and for semiconsolidated, saturated media, it is 36° C/km (David S. Chapman, University of Utah, oral commun., 1986).

Assuming an average surface temperature of 6° C (the mean annual air temperature of the study area) and using average geothermal gradients along the Wasatch Front, temperatures of bedrock at specific depths can be computed. Assuming that thermal conductivities are uniform, assuming that groundwater flow does not significantly disturb conductive geothermal gradients, and assuming the groundwater has sufficient time to equilibrate with the temperature of the bedrock with which it is in contact, the depth of groundwater circulation can be approximated. Based on these assumptions, estimated ranges of depths of circulation of groundwater supplying the springs examined for the study are presented in table 13. Because it is not clear whether or not groundwater flow disturbs conductive geothermal gradients in the study area, the depths reported in table 13 suggest that groundwater in the study area circulates at relatively shallow depths as opposed to depths of one to two km. However, the fluctuations in groundwater temperatures suggest that

Table 13. Estimated depths of groundwater circulation

Spring	Geothermal Gradient[a] °C/km	Range of Groundwater Temperature °C	Range of Depths of Circulation m
Peter Jensen	30	10.2-12.8	140-225
West Hallings	30	11.5-18.9	185-430
Mud	30	9.4-12.7	115-225
Mantua	18	5.6-12.7	10-370
Maple	30	6.6-10.1	20-135

[a]gradient listed for the type of material from which each spring issues

the springs may be supplied by shallow as well as deeper sources of groundwater.

The data in table 13 suggest that Peter Jensen and Mud Springs have similar depths of circulation. The narrow ranges of temperature fluctuations suggest that the two springs may be supplied by a number of shallow sources of groundwater as opposed to both shallow and deep sources.

Temperature data for West Hallings Spring suggest that groundwater supplying it has the greatest depth of circulation of all the Mantua Valley springs studied (table 13). The Wellsville Mountains, topographically higher and northwest of West Hallings Spring, have sufficient relief to allow groundwater to circulate at relatively great depths. The wide range of temperature fluctuations suggests that West Hallings Spring may sometimes be supplied by a deeper source of groundwater than some of the other springs.

Based on the temperature data for Mantua Spring, groundwater depths of circulation range from 10 m to 370 m. This range of circulation depths is wider than that of any of the other springs (table 13), and it may be a function of sampling procedures. A northeast-trending lineament mapped near Mantua Spring (pl. 2) suggests that a zone of higher permeability may exist at depth. If such a wide range of groundwater circulation depths for Mantua Spring actually exists, it may be the result of shallow groundwater (cooler) mixing with deeper groundwater (warmer) along a zone of permeability.

Temperature data for Maple Spring suggest that depth of circulation of groundwater supplying the spring is relatively shallow (table 13). Saturation indices of groundwater from Maple Spring (fig. 8) show that the water was not saturated with respect to calcite. Bedrock located between Maple Spring and its probable recharge area is predominantly limestone and dolostone, so calcite is clearly available for solution. The absence of calcite-saturated water at Maple Spring might indicate that the flow-through time is relatively rapid so that groundwater would not have time to equilibrate with warmer bedrock at depth. The temperature data then could be misleading, and the depth of circulation could be deeper than indicated.

Groundwater Flow Paths

Langmuir (1971) reported that molar Ca^{+2}/Mg^{+2} ratios in groundwater have both hydrologic and geochemical implications. In his study of carbonate aquifers in central Pennsylvania, Langmuir (1971) found that: 1) groundwater which flows through mapped limestones had Ca^{+2}/Mg^{+2} ratios greater than or equal to 2.2; 2) Ca^{+2}/Mg^{+2} ratios lower than 2.2 suggest the presence of dolomitic beds in limestone formations or groundwater that originated in adjacent dolostones; 3) cross-formational groundwater flow, as in the preceding statement, may occur along zones of rock fracture which have been enlarged by solution; and 4) Ca^{+2}/Mg^{+2} ratios in mapped dolostone areas are less than or equal to 1.5.

As $CaMg(CO_3)_2$ dissolves, Ca^{+2} and Mg^{+2} ions are released in equal amounts, so that milliequivalents of Ca^{+2} and Mg^{+2} are equal (Hem, 1970, p. 143). Groundwater undersaturated or saturated with respect to dolomite then, flowing through pure dolostones, would have a Ca^{+2}/Mg^{+2} ratio equal to 1.0, assuming that congruent dissolution of dolomite occurred. To achieve Ca^{+2}/Mg^{+2} ratios of 1.5, as in Langmuir's (1971) dolostones, the dolostones being dissolved must be: 1) impure; or 2) undergoing incongruent dissolution.

Berner (1978) reported that the dissolution rate of minerals at a fixed degree of undersaturation is affected by one of the following processes: 1) transport-control; 2) surface reaction-control; or 3) mixed transport and surface reaction-control. In transport-controlled dissolution, ions are detached from the crystal so rapidly that they cause the solution to be saturated near the surface of the crystal; the rate of dissolution is dependant on flow velocity. In surface reaction-controlled dissolution, ions are detached from the crystal surface at a slow enough rate that advection and diffusion cause the concentration of solute to be the same at the surface of the crystal as that in the surrounding solution; increased flow has no effect on the rate of dissolution. In mixed transport and surface reaction-controlled dissolution, concentrations of solute are greater at the crystal surface than in the surrounding solution but lower than that necessary for saturation. Berner (1978) indicated that the solubility of $CaCO_3$ is 6 X 10^{-5} mole/liter (solubility product = 3.6 X 10^{-9}) and that its dissolution is controlled by surface reaction processes. The solubility product of $CaMg(CO_3)_2$ is approximately 10^{-17} (Drever, 1982). Because the solubility of dolomite is much less than that of calcite, the dissolution of dolomite is probably also controlled by surface reaction processes. Because dissolution of calcite and dolomite are controlled by surface reaction processes, sudden influxes of water, such as recharge to a groundwater system by snowmelt, should not affect the rates of dissolution of the two minerals. However, given a constant degree of undersaturation, the amount of calcite dissolved in limestone may be greater than the amount of dolomite dissolved in dolostone because of the different solubilities of the two minerals. Ca^{+2}/Mg^{+2} ratios should be used with caution when attempting to qualitatively describe the relative amounts of limestone and dolostone through which groundwater has flowed.

Molar Ca^{+2}/Mg^{+2} ratios of Mantua Valley carbonate springs and the well are presented in table 14. Based on Langmuir's (1971) criteria and the Ca^{+2}/Mg^{+2} ratios in table 14, Mantua Valley carbonate springs should be flowing through impure dolostones, dolostones undergoing incongruent dissolution, or possibly mixed limestone and dolostone terrane.

Peter Jensen Spring has the highest average Ca^{+2}/Mg^{+2} ratio (1.7) of the Mantua Valley springs, which suggests that the groundwater flows through both limestone and dolostone, and possibly through more limestone than dolostone lithologies. Plate 1 shows that mapped units topographically above Peter Jensen Spring include the Mississippian Deseret

Table 14.--Ca^{+2}/Mg^{+2} molar ratios of Mantua Valley carbonate springs and well (n.d., no data)

	P. Jensen	W. Hallings	Mud	Maple	Well
Mar	1.8	1.4	1.5	1.6	1.7
May	1.6	1.2	n.d.	1.4	1.3
Jun	1.6	1.2	n.d.	1.5	1.4
Jul	1.6	1.2	1.3	1.4	1.4
Sep	1.8	1.5	1.6	1.7	1.8
Nov	1.8	1.5	1.5	1.8	1.7
Jan	1.7	1.4	1.4	1.7	1.6
Mar	1.6	1.4	1.4	1.8	1.5
May	1.7	1.4	1.4	1.9	1.5
$\bar{x}$	1.7	1.4	1.4	1.6	1.5

Limestone and the Humbug and Brazer Formations. These Mississippian units comprise limestones, dolostones, chert, and quartzitic sandstone (Crittenden and Sorensen, 1985). For water to recharge on the hillslope to the east of Mantua Reservoir and discharge from the springs at the base of the hill, the groundwater would have to have cross-formational flow, since bedrock strikes northwest in that area. Plate 2 shows that N. 20° E. dominant joints and northeast-trending lineaments are located in the general area of Peter Jensen Spring. The N. 20° E. joint set and/or the lineament probably provide sufficient permeability for groundwater to flow to Peter Jensen Spring.

Both West Hallings and Mud Springs have average Ca^{+2}/Mg^{+2} ratios of 1.4, suggesting that flow is predominantly through dolostone. West Hallings Spring discharges from the Cambrian Bloomington Formation (pl. 1). The basal member of the Bloomington Formation, the Hodges Shale (siltstone, mudstone, and shale with interbeds of limestone; Crittenden and Sorensen, 1985), probably acts as an aquitard, concentrating flow within the Unnamed Limestone Member or limestone interbeds of the Calls Fort Shale Member. The Unnamed Limestone Member comprises 1 to 3 m thick, laminated dolostone beds (Crittenden and Sorensen, 1985), through which groundwater may flow. The low Ca^{+2}/Mg^{+2} ratios characteristic of West Hallings Spring may be, in part, acquired from the laminated dolostone beds. Plate 2 shows dominant northwest-striking joints to the north of West Hallings Spring, suggesting that recharge water from the east flank of the Wellsville Mountains could flow along the northwest-trending joints and be concentrated in the Bloomington Formation to discharge as West Hallings Spring.

Mud Spring discharges from the Cambrian St. Charles Dolostone, close to its contact with the Ordovician Garden City Formation. Both formations contain limestone, dolostone, and chert. Some of the water which recharges to the north of Mud Spring could flow along the St. Charles-Garden City contact (a dolostone-limestone stratigraphic contact). Joint orientations plotted on plate 2 topographically above Mud Spring are dominantly N. 20° E., suggesting that water which recharges northeast of Mud Spring may flow across strike to Mud Spring. This cross-formational flow, through mixed dolostone and limestone units, may also contribute to the low Ca^{+2}/Mg^{+2} ratios of Mud Spring.

Fracture orientations topographically above Mantua Spring, shown on plate 2, suggest that groundwater could flow along several joint sets. A lineament shown on plate 2 indicates that a permeable zone at depth may act as an avenue for groundwater flow. The lack of abundant, massive outcrops above Mantua Spring suggests that bedrock is highly fractured in that area. Groundwater could flow along a number of narrow joints in the quartzites and be concentrated along the lineament to discharge as Mantua Spring.

Ca^{+2}/Mg^{+2} ratios for Maple Spring (average = 1.6) suggest that the groundwater flows through a mixed carbonate terrane, but like Peter Jensen Spring, the presence of more limestone along the flow path may account for the slightly higher ratio. Maple Spring discharges from the Calls Fort Shale Member of the Bloomington Formation, which comprises shale interbedded with limestone and silty limestone (Crittenden and Sorensen, 1985). One or more shale beds may act as confining layers, forcing groundwater flow through the limestone interbeds. No Bloomington bedrock is exposed between Maple Spring and Sink Hole Valley, which is topographically higher than the spring. Groundwater supplying Maple Spring must have a cross-formational flow path along joints widened by solution if it recharges at Sink Hole Valley and discharges at Maple Spring. Dominant joints near Sink Hole Valley, plotted on plate 2, trend northwest. Groundwater could flow from Sink Hole Valley along northwest-striking joints until it reaches Bloomington Formation bedrock, where it may flow along strike to discharge as Maple Spring.

Conceptual Models

Conceptual models of carbonate aquifers have been presented in Previous Investigations. Shuster and White (1971) classified springs in the Central Appalachians as diffuse flow and conduit flow types, based on hydrogeologic evidence. By comparing chemical parameters of the two types of springs, Shuster and White (1971) found that hardness coefficients of variation of diffuse flow springs were generally less than 5%, while hardness variations of conduit springs ranged from 10% to 24%. Todd (1980, p. 282) defined $CaCO_3$ hardness (total hardness), as $H_T = 2.5\ Ca^{+2} + 4.1\ Mg^{+2}$, where H_T, Ca^{+2}, and Mg^{+2} are in milligrams per liter. Shuster and White (1971) assumed that total hardness is a significant chemical variable of carbonate springs and used the coefficient of variation, $CV = \sigma/\bar{x} \times 100$, where σ is the standard deviation and $\bar{x}$ is the arithmetic mean of total hardness. Quinlan and Ewers (1981, p. 490) prefered to use the specific conductivity coefficient of variation to distinguish between spring types. Specific electrical conductance is a general indication of dissolved solids (Todd, 1980, p. 281). Since specific conductivity measurements were not taken for this study, coefficients of variation for total dissolved solids were computed instead. Chemical parameters used to classify the carbonate aquifers which supply Mantua Valley springs are presented in table 15.

Recharge Areas and Water Balance

Recharge areas for the Mantua Valley springs were estimated using a graph presented by Todd (1980, fig. 2.16, p. 49), which shows the logarithmic relationship between recharge area, annual recharge, and average discharge of a spring. Values for average discharge of each spring were assumed to be close to the single-measurements presented in table 1, because substantial seasonal fluctuations in discharges of the springs were not observed. Most of the recharge to the groundwater systems occurs from melting of the winter snowpack at the higher elevations. Bjorklund and McGreevy (1974) calculated the recharge to groundwater in the lower Bear River drainage basin by evaluating the sources and quan-

Table 15.--Summary of chemical parameters of carbonate springs and well

Station	Average Total Hardness as $CaCO_3$ (mg/l)	Hardness Coefficient of Variation (%)	Average Total Dissolved Solids (mg/l)	Total Dissolved Solids Coefficient of Variation (%)	Average CO_2 Pressure (atm)	Average Calcite Saturation Index
Peter Jensen	211	5.1	256	3.0	$10^{-1.78}$	-0.372
West Hallings	192	6.9	247	2.8	$10^{-1.93}$	-0.286
Mud	208	6.8	251	3.1	$10^{-1.89}$	-0.230
Maple	157	9.1	193	5.8	$10^{-1.97}$	-0.453
Well	236	6.0	309	3.5	$10^{-1.83}$	-0.059

tities of water available for recharge. Calculations for recharge to the Mantua Valley and Sink Hole Valley watersheds are presented in table 16. Values for recharge and precipitation on recharge areas reported by Bjorklund and McGreevy (1974, p. 16) compare favorably with the values calculated for the Mantua Valley study area, i.e., recharge is approximately 57% of the annual precipitation for the lower Bear River drainage basin. Values for annual recharge to the springs in the Mantua Valley study area were calculated to be approximately 49% of the annual precipitation values reported by Bjorklund and McGreevy (1974, fig. 11, p. 47).

Estimated recharge areas of the Mantua Valley springs are shown in table 17. Recharge areas of the springs were drawn on topographic maps taking into account the following: 1) the assumption that recharge areas are topographically higher than the springs; 2) hydrostratigraphic units and the chemistry of water from each spring; and 3) lineaments topographically above the springs which imply zones of permeability at depth and which should be included in the recharge areas. The recharge areas were drawn to scale by overlaying transparent pieces of paper representing the recharge areas reported in table 17. Figure 9 shows the relationships of the recharge areas of the Mantua Valley springs.

Approximately 73,862 m^3/day of water (21,866 ac-ft/yr; Bjorklund and McGreevy, 1973; on file with Utah State Dept. of Health) discharges as springs in Mantua Valley. If 77,219 m^3/day (22,860 ac-ft/yr) of water recharges the groundwater system (table 16) and 73,862 m^3/day discharges as springs, the difference, 3357 m^3/day (994 ac-ft/yr), may either: 1) discharge in Mantua Valley as ungagued diffuse seepage; 2) flow through fractured quartzites westward to Salt Lake Valley; 3) contribute to

Table 16.--Recharge calculations for Mantua Valley and Sink Hole Valley watersheds (91 km^2)

	m^3/day	ac-ft/yr
Water available for recharge		
Precipitation on recharge areas[a]	159,114	47,104
Subsurface inflow[b]	676	200
Total	159,790	47,304
Water unavailable for recharge		
Evapotranspiration in recharge areas[c]	77,909	23,064
Runoff from recharge areas[d]	4662	1380
Total	82,571	24,444
Recharge (water available less evapotranspiration and runoff)	77,219	22,860

Recharge as % of precipitation = (77,219 ÷ 159,114) X 100 = 49%

[a]based on 686 mm/yr (Bjorklund and McGreevy, 1974, fig. 11, p. 47) over a recharge area of 83 km^2

[b]estimate

[c]average annual evapotranspiration at elevations of 1524-1829 m (5000-6000 ft) = 345 mm (13.59 in.); 1829-2134 m (6000-7000 ft) = 348 mm (13.72 in.); 2134-2438 m (7000-8000 ft) = 316 mm (12.43 in.; Bjorklund and McGreevy, 1974)

[d]Box Elder Creek, average of four year record (Price and Jensen, 1982)

Table 17.--Estimated recharge areas of Mantua Valley springs

Spring	Annual Recharge[a]		Spring Discharge[b]	Recharge Area[c]	
	mm	in.	l/s	km^2	mi^2
Peter Jensen	311	12.2	31	3.0	1.2
West Hallings	311	12.2	227	18.0	6.9
Mud	311	12.2	99	9.0	3.5
Mantua	311	12.2	3	0.3	0.1
Maple	373	14.7	90	7.0	2.7

[a]49% of annual precipitation value reported by Bjorklund and McGreevy (1974, fig. 11, p. 47)
[b]from table 1
[c]estimated from Todd (1980, fig. 2.16, p. 49)

groundwater storage; or 4) be used in any combination of the above.

Peter Jensen Spring

Peter Jensen Spring is thought to issue from a diffuse flow aquifer because: 1) it has low hardness and TDS coefficients of variation (table 15); 2) it has the highest average total dissolved solids concentration (table 15); 3) its rate of discharge is relatively low (table 1); 4) its rate of discharge does not seem to be highly variable (Kenny Charlie, Brigham City Corp., oral commun., 1986); 5) field evidence suggests that the spring discharges in an area of many small springs and seeps; and 6) other springs on the east side of Mantua Valley, which probably flow through the same aquifer, also discharge small quantities of water in low-lying, marshy areas.

Although groundwater from Peter Jensen Spring has a relatively low

Figure 9.--Recharge areas of Mantua Valley Springs. ---, possible recharge areas, which correspond with the values shown in table 18 for each spring; —, boundary of study area. Lines of section for conceptual models are labeled. Contours in feet; 1 ft = 0.3048 m.

average calcite saturation index (table 15), it did reach saturation with respect to calcite in November, 1985 (fig. 8). All of the carbonate springs reached saturation with respect to calcite in November, except Maple Spring, which came closer to saturation in November than in any other month (fig. 8). This coincidence of saturation in November suggests that shallow flow components of the systems had subsided and deeper flow components were dominating discharge. The absence of major recharge events during the summer and fall months preceding November may have caused the shallow flow components to diminish or disappear. Although diffuse flow springs examined by Shuster and White (1971) are generally near saturation with respect to calcite, two of their diffuse flow springs were undersaturated on average.

The recharge area for Peter Jensen Spring is estimated to be 3.0 km^2 (1.2 mi^2) (table 17). Because the relatively high Ca^{+2}/Mg^{+2} ratios of the groundwater, the N. 20° E. joint sets, and the lineaments topographically above the spring suggest that the groundwater is flowing through the Upper Paleozoic carbonate aquifer (pl. 1), the recharge area for Peter Jensen Spring is thought to be the area to the east of the spring outlet. The asymmetric topography to the east of Peter Jensen Spring suggests that the groundwater divide may be offset to the east of the surface-water divide, therefore the recharge area for Peter Jensen Spring is shown to extend east of the surface-water divide (fig. 9).

Shuster and White (1971) suggested that CO_2 pressures are more strongly controlled by source areas of recharge than by aquifer flow characteristics. Drever (1982, p. 55) stated that CO_2 derived from the atmosphere can be increased by percolation of recharge water through soil. This additional CO_2 increases the aggressiveness of the water and thus the amount of $CaCO_3$ that the water can dissolve (Drever, 1982, p. 55). The fact that the average CO_2 pressure of groundwater from Peter Jensen Spring is slightly higher than that of the other carbonate springs (table 15), suggests that Peter Jensen Spring recharge may be percolating through more soil than recharge to other springs. A thin regolith cover occurs on the steep, west-facing slope of the upper portion of the Peter Jensen Spring recharge area, as evidenced by the numerous outcrops in that area. However, some of the recharge probably occurs on the topographically lower, gentler slope above the spring. This lower area appears to be mantled by thicker colluvial and slope-wash deposits, through which some recharge water may percolate, increasing its CO_2 content. This additional CO_2 in Peter Jensen Spring groundwater might be responsible for its relatively low average saturation index (table 15). A conceptual model of Peter Jensen Spring showing its relationship to a possible larger-scale flow system is presented on figure 10.

West Hallings and Mud Springs

West Hallings and Mud Springs have higher hardness coefficients of variation, similar TDS coefficients of variation, and are on average less undersaturated with respect to calcite than is Peter Jensen Spring (table 15). West Hallings, Mud, and Peter Jensen Springs all have different magnitudes of discharge, i.e., 227 l/s, 99 l/s, and 31 l/s respectively (table 1). These differences in chemical parameters and magnitudes of discharge suggest that West Hallings and Mud Springs issue from aquifer types different from the aquifer supplying Peter Jensen Spring. Although the hardness coefficients of variation of West Hal-

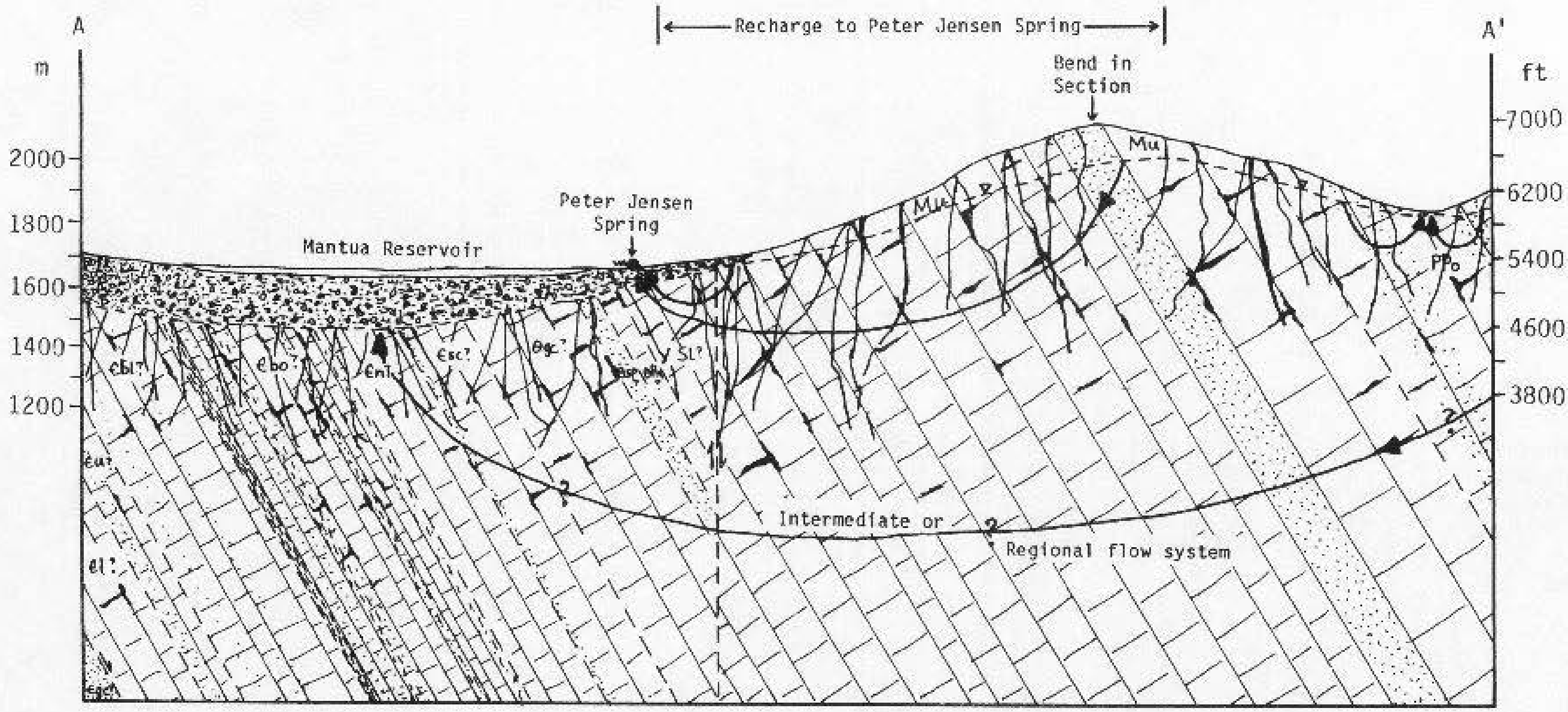


Figure 10.--Conceptual model of Peter Jensen Spring and its relationship to regional groundwater flow. €gc, Geertson Canyon Quartzite; €l, Langston Dolostone; €u, Ute Limestone; €bl, Blacksmith Dolostone; €bo, Bloomington Formation; €n, Nounan Dolostone; €sc, St. Charles Dolostone; θgc, Garden City Formation; θsp, Swan Peak Formation; θfh, Fish Haven Dolostone; Sl, Laketown Dolostone; Mu, Mississippian rocks, undifferentiated; PPo, Oquirrh Formation; Qcs, colluvium and slopewash, includes some alluvium. Geology adapted from Dover (1985). No vertical exaggeration.

lings and Mud Springs are higher than that of Peter Jensen Spring, the higher values are not of sufficient magnitude to warrant classifying the aquifers as conduit flow types, based on the criteria presented by Shuster and White (1971). Wilson (1979), after personal communication with White, characterized two springs with coefficients of variation of 7.2% and 6.0% as flowing through an open fracture system, as opposed to a conduit system. West Hallings and Mud Springs might be best characterized as flowing through open fracture systems, since they do not display characteristics of true conduit flow nor true diffuse flow aquifers.

The tritium activity of 16.02 TU in one sample from West Hallings Spring is difficult to interpret. One tritium analysis can be very ambiguous, whereas monthly analyses over a period of a year might yield a sequence of tritium data which could be matched with a previous sequence of atmospheric tritium data adjusted for radioactive decay. However, even a sequence of tritium data can give ambiguous results because of the possibility of mixing of older and younger waters. The sequence of tritium data weighted for precipitation and adjusted for radioactive decay is presented on figure 11. Three tritium activities, which have been weighted for precipitation and adjusted for radioactive decay (fig. 11; appendix D), are close ($\pm$0.5 TU) to the value of the tritium activity of groundwater from West Hallings Spring (table 12). All three of the values correspond to recharge water which was less than nine years old in June 1986, i.e., 8.9 yr, 6.25 yr, and 3.2 yr. West Hallings Spring groundwater is older than six months, since atmospheric tritium in precipitation which fell in the area during the winter of 1985-86 had an activity of 8.86 TU in June, 1986 (table 12). West Hallings Spring

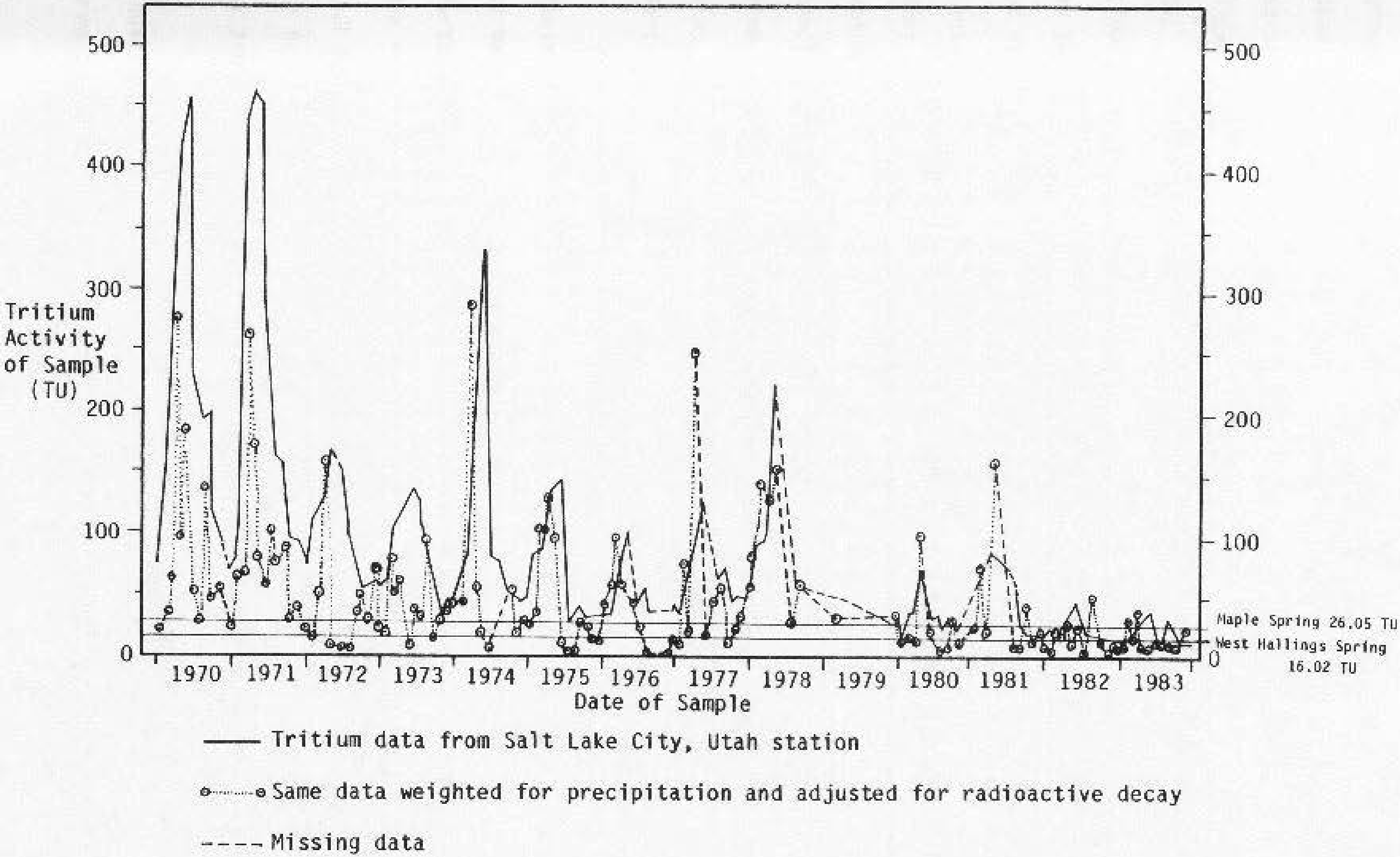


Figure 11.--Monthly tritium data and data weighted with precipitation and adjusted for radioactive decay, 1970-83, Salt Lake City, Utah.

groundwater is thought to have a mean residence time nearer to 8.9 years than to 3.2 years for the following reasons: 1) the groundwater was saturated with respect to calcite in November (presumably during baseflow) and again in May (presumably just prior to peak discharge) (fig. 8); 2) the depth of groundwater circulation is estimated to be relatively deep, i.e., greater than 185 m (table 13); and 3) water recharging on the east flank of the Wellsville Mountains may require a long period of time to percolate along narrow joints and/or bedding planes in the Langston, Ute, and Blacksmith Formations before reaching joints and bedding planes widened by solution in the Bloomington Formation, through which it may travel more quickly. Because of the lack of data, definite conclusions cannot be made; however, the tentative conclusion derived from the tritium data, based on the assumption that simple decay and plug flow have occurred, is that the mean residence time of West Hallings Spring groundwater is slightly less than nine years.

West Hallings Spring is located along an inferred northeast-trending fault (pl. 1). The recharge area for West Hallings Spring is inferred to be the east flank of the Wellsville Mountains because of the dominant northwest-striking joints in that area (pl. 2). Table 17 indicates that the recharge area for West Hallings Spring is approximately 18.0 km^2 (6.9 mi^2). The relationship of the West Hallings Spring recharge area to the recharge areas of the other springs is shown on figure 9. Figure 12 shows a conceptual model of West Hallings Spring.

Mud Spring is located along the same inferred northeast-trending fault along which West Hallings Spring is located (pl. 1). The recharge area for Mud Spring is inferred to be the area north and northeast of

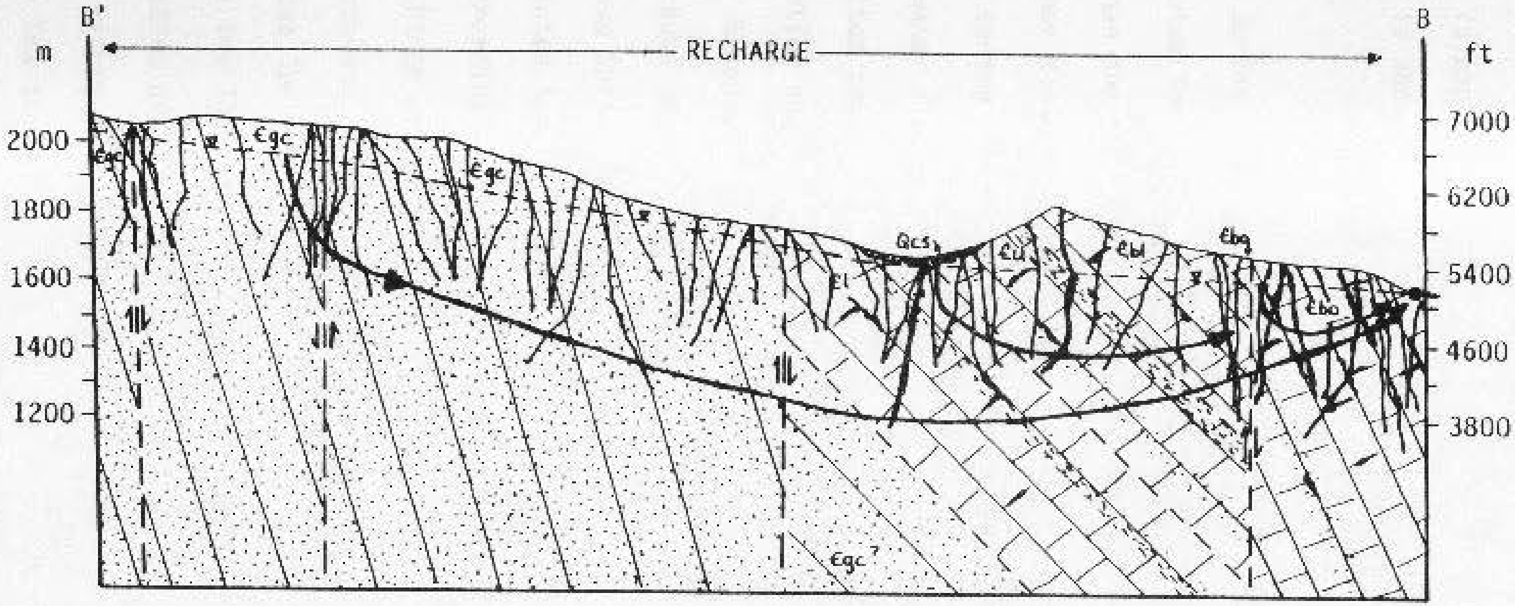


Figure 12.--West Hallings Spring conceptual model. €gc, Geertson Canyon Quartzite; €l, Langston Dolostone; €u, Ute Limestone; €bl, Blacksmith Dolostone; €bo, Bloomington Formation; Qcs, colluvium and slopewash, includes some alluvium. Geology adapted from Dover (1985). No vertical exaggeration.

the spring outlet because of the dominant northeast-trending joints in that area (pl. 2). Table 17 indicates that the recharge area for Mud Spring is approximately 9.0 km^2 (3.5 mi^2). The recharge area of Mud Spring is shown on figure 9. Figure 13 shows a conceptual model of Mud Spring.

Mantua Spring

Mantua Spring most certainly has different controls on groundwater flow than the carbonate springs, since the groundwater neither issues from, nor flows through, carbonate rocks. The very low discharge of Mantua Spring (table 1) suggests that the quartzites through which groundwater flows have low permeabilities. Low permeability of the quartzites should be expected, since the joints have not been affected by solution and fracture widths were observed to decrease rapidly with depth. Because of the steep slopes topographically above Mantua Spring and because of the low permeability of the quartzites, the recharge area of Mantua Spring is probably much larger than the area reported in table 17. Mantua Spring is located near the intersection of an inferred north-trending fault (pl. 1) and a northeast-trending lineament (pl. 2). The recharge area of Mantua Spring probably includes the lineament to the southwest of the spring shown on plate 2. A possible recharge area of Mantua Spring is shown on figure 9.

The low TDS concentration in groundwater from Mantua Spring may be a combination of a short residence time of the groundwater and the unreactive nature of the quartzites through which the water flows. The relatively low temperature and pH of the groundwater from Mantua Spring tend to retard dissolution of silica.

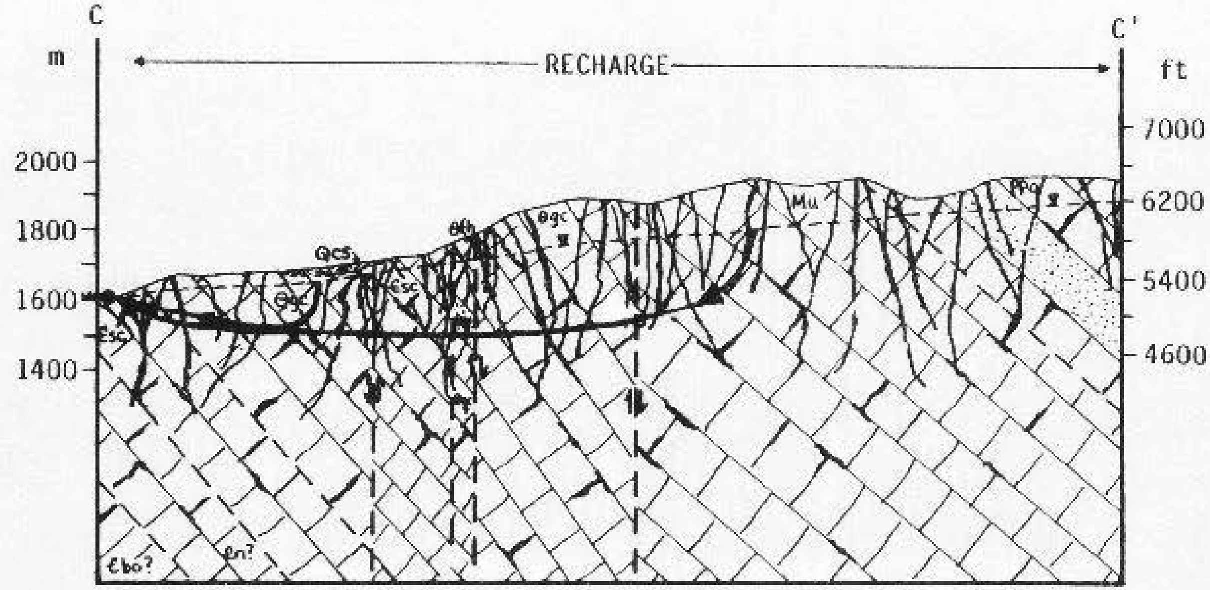


Figure 13.--Mud Spring conceptual model. €bo, Bloomington Formation; €n, Nounan Dolostone; €sc, St. Charles Dolostone; θgc, Garden City Formation; θsp, Swan Peak Formation; θfh, Fish Haven Dolostone; Mu, Mississippian rocks undifferentiated; PPo, Oquirrh Formation; Qcs, colluvium and slopewash, includes some alluvium. Geology adapted from Dover (1985). No vertical exaggeration.

The turbidity which sometimes appears in Mantua Spring may be related to recharge water percolating through the soil zone, where it might acquire clay particles, identified through X-ray diffraction, and/or micro-organic particles. If Mantua Spring is a local flow system with a relatively short residence time, the groundwater would not have sufficient time to dissolve the clays before they are discharged at the spring orifice. A conceptual model of Mantua Spring is presented on figure 14.

Maple Spring

The hardness and TDS coefficients of variation of Maple Spring are the highest values of the Mantua Valley carbonate springs (table 15). Data suggesting that Maple Spring might be controlled by a conduit flow system are:

1) The relatively high hardness coefficient of variation of 9.1%. Shuster and White (1971) classified a spring with the same hardness coefficient of variation as controlled by a conduit flow system.

2) Maple Spring groundwater was not observed to be saturated with respect to calcite.

3) Three sinkholes are located topographically above Maple Spring.

4) Maple Spring groundwater becomes turbid during high flow and after heavy rains.

5) Maple Spring groundwater has a relatively low TDS content, suggesting rapid flow.

6) Maple Spring groundwater has a relatively low temperature, suggesting shallow circulation or deeper, but rapid, flow.

Several lines of evidence, however, suggest that Maple Spring may

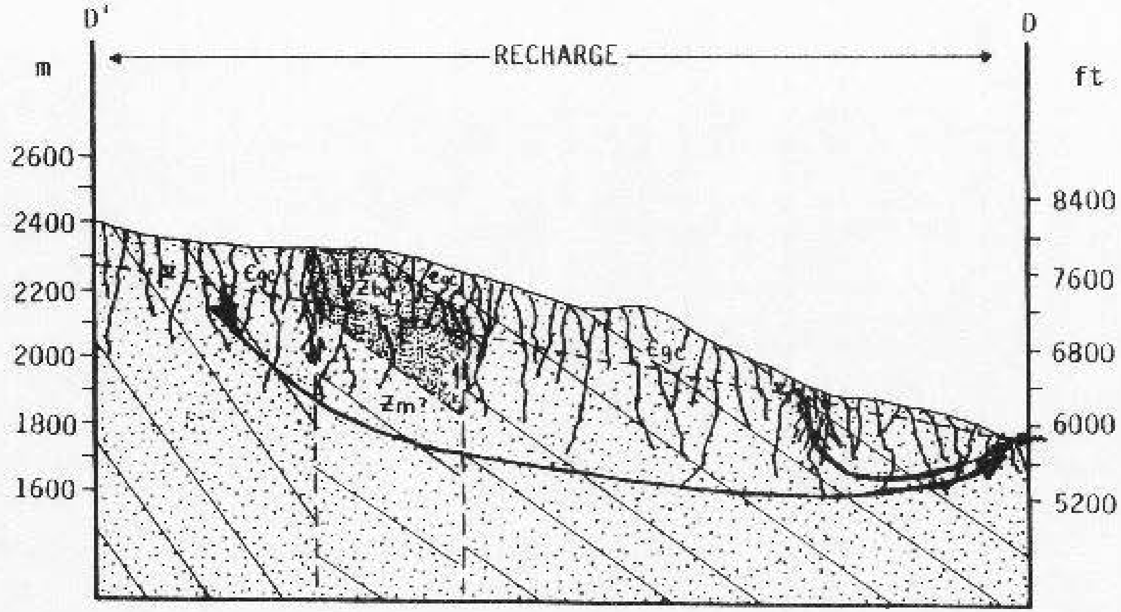


Figure 14.--Mantua Spring conceptual model. Zm, Mutual Formation; Zbq, Browns Hole Formation; Ꞓgc, Geertson Canyon Quartzite. Geology adapted from Crittenden and Sorensen (1985). No vertical exaggeration.

be controlled by a system in transition between a diffuse flow and a conduit flow system:

1) Shuster and White (1971) interpreted the system controlling flow to Weaver Spring (hardness coefficient of variation = 10.23%) as in a transition state between a diffuse flow and a conduit flow system, because the chemical parameters suggest conduit flow but only a few shallow sinks are located topographically above the spring.

2) A different spring classified by Shuster and White (1971) as controlled by a diffuse flow system has a few shallow sinks associated with it.

3) The slope of the recession curve (Waddell and others, 1986) of the 1985 discharge of Maple Spring (greater than 365 days to recede one log cycle) suggests that the seasonal variability of discharge of Maple Spring is very small. The large discharge and the small seasonal variability of Maple Spring suggest that the aquifer supplying Maple Spring has a large storage capacity and a relatively large transmissivity (Waddell and others, 1986).

4) Maple Spring actually discharges from a bedrock outlet as well as a seep area approximately 30 m (100 ft) away. Chemical analyses performed by Ford Chemical Laboratory, Inc., Salt Lake City, Utah in 1971 (on file at Mantua Fish Hatchery) show that groundwater from Upper (bedrock outlet) and Lower (seep area) Maple Springs are essentially the same, with Upper Maple Spring TDS only 6 mg/l greater than that of Lower Maple Spring. This chemical similarity of groundwater from the two spring outlets suggests that the water originates from the same source and/or has the same or very similar flow paths. However, a conduit flow system

usually discharges through a single large spring (Shuster and White, 1971).

5) All of the springs classified by Shuster and White (1971) as controlled by conduit flow systems were sometimes or usually turbid. However, Weaver Spring, the spring which Shuster and White (1971) interpreted as being in a transition state, became turbid during exceptionally high flows, and a spring classified as controlled by a diffuse flow system became turbid after heavy rains, so turbidity is not necessarily an absolute indicator of a conduit flow system.

6) The sinkholes located topographically above Maple Spring may be, in part, the result of tectonic activity and not totally the result of karst processes. No other sinkholes were identified between Sink Hole Valley and the spring outlet.

7) Some springs identified by Shuster and White (1971) as being controlled by diffuse flow systems discharge water which was on average undersaturated with respect to calcite.

Obviously there is a continuum between ideal diffuse flow and conduit flow systems and objectively differentiating between the two types of systems is not always possible or desirable. Maple Spring is thought to be controlled by an open-fracture system, more open than the systems controlling flow to West Hallings and Mud Springs, but not as open as a conduit system, for the following reasons:

1) Maple Spring displays characteristics which suggest a system in transition between a diffuse flow and a conduit flow system.

2) Karst features described by Shuster and White (1971), as indicative of conduit flow systems, such as lines of sinkholes above springs, sink-

ing streams, and caves with solution tubes and streams, are not present in the study area.

3) The high hardness and TDS coefficients of variation of Maple Spring suggest that the fracture system controlling flow to Maple Spring might be more open than those of West Hallings and Mud Springs. In addition, Mud and West Hallings Springs reached saturation with respect to calcite, whereas Maple Spring did not.

During periods of low flow, and when there have been no recent recharge events (snowmelt or heavy rains), the groundwater flowing from Maple Spring is non-turbid. Maple Spring becomes turbid during the spring while the snow is melting and as quickly as three days after a heavy rain (Ron Roubidoux, oral commun., 1986). The appearance of turbidity after recharge events suggests that one or more shallow, local flow systems may mix with Maple Spring groundwater flowing through the open-fracture system. The recharge water may percolate through the regolith, picking up clays and organic material, which constitute the turbidity. Alternatively, the turbidity could originate in the main recharge area and flow through the entire system. However, for the turbidity to travel from the main recharge area to the outlet within three days, without being filtered, would require a very open conduit flow system, which is not evident for reasons previously cited. If the system were open enough to allow water to flow at such high velocities, larger pieces of organic material such as leaves and twigs might be expected to be discharged from the spring.

The tritium activity for Maple Spring, like that of West Hallings Spring, is difficult to interpret. Again, the preliminary nature of the following interpretation must be emphasized, because only one sample was analyzed and there is the possibility that mixing of waters has occurred. However, because the sample was taken several weeks after any major precipitation events had occurred, it is hoped that any groundwater contributed by a local flow system had previously discharged. Figure 11 and appendix D show that, of the tritium data which have been weighted for precipitation and adjusted for radioactive decay, four samples had activities in June 1986 which were close (±0.5 TU) to the value of the tritium activity of groundwater from Maple Spring (table 12). The four values suggest that the mean residence time of Maple Spring groundwater may be 12.6 yr, 11.5 yr, 10.7 yr, or 4.1 yr. Like West Hallings Spring, Maple Spring groundwater is older than six months, since precipitation which fell in the area during the winter of 1985-86 had a tritium activity of 8.86 TU in June, 1986.

The mean residence time of Maple Spring groundwater is thought to be closer to 4.1 yr than to 10.7 to 12.6 yr because the groundwater was not observed to be saturated with respect to calcite. Explanations for undersaturated groundwater with respect to calcite are as follows: 1) no calcite is available along the flow path; 2) mixing of two waters saturated with respect to calcite results in an undersaturated water (Bloom, 1978, fig. 7-3, p. 141); and 3) the flow path is very short and/or the flow-through time is rapid. In the case of Maple Spring, all of the options except for (3) can be refuted: 1) the spring discharges from a limestone unit and presumably flows through several other mixed carbonate formations; and 2) recharge water which might mix with saturated groundwater already in the system would most likely not be saturated, so the mixing of two saturated waters would not occur. Of the possibilities suggested, the best explanation for the calcite-undersaturated nature of Maple Spring groundwater is that it flows along a short flow path and/or the travel time is rapid enough that the water does not have time to reach equilibrium with respect to calcite. Other evidence suggesting that the mean residence time of Maple Spring groundwater is relatively short is the high correlation between annual precipitation cumulative departure from normal and Maple Spring annual discharge departure from normal (fig. 15). Figure 15 shows that a drier (or wetter) than normal year seems to cause the spring discharge to decrease (or increase) approximately four years later. Although a longer record would be more desirable, the data in figure 15 support the interpretation of the tritium data.

Maple Spring is not located along any mapped faults (pl. 1) or lineaments (pl. 2), although dominant northwest-trending joints are located southeast of the spring (pl. 2). The recharge area for Upper Maple Spring is estimated to be 7.0 km^2 (2.7 mi^2) (table 17), and for both Upper and Lower Maple Springs combined, the recharge area is approximately 13.0 km^2 (5.0 mi^2). The recharge area for Maple Spring probably includes Sink Hole Valley, since it is topographically higher than the spring outlet and contains three sinkholes at its lowest point. Mantua residents report that during rapid snowmelt in the spring, water from the stream draining Sink Hole Valley pours down into the sinkholes in the southwest corner of the valley. The recharge area for the Maple Springs is shown on figure 9. If Maple Spring groundwater recharges at the sinkholes in Sink Hole Valley, it has a maximum velocity of 2 m/day

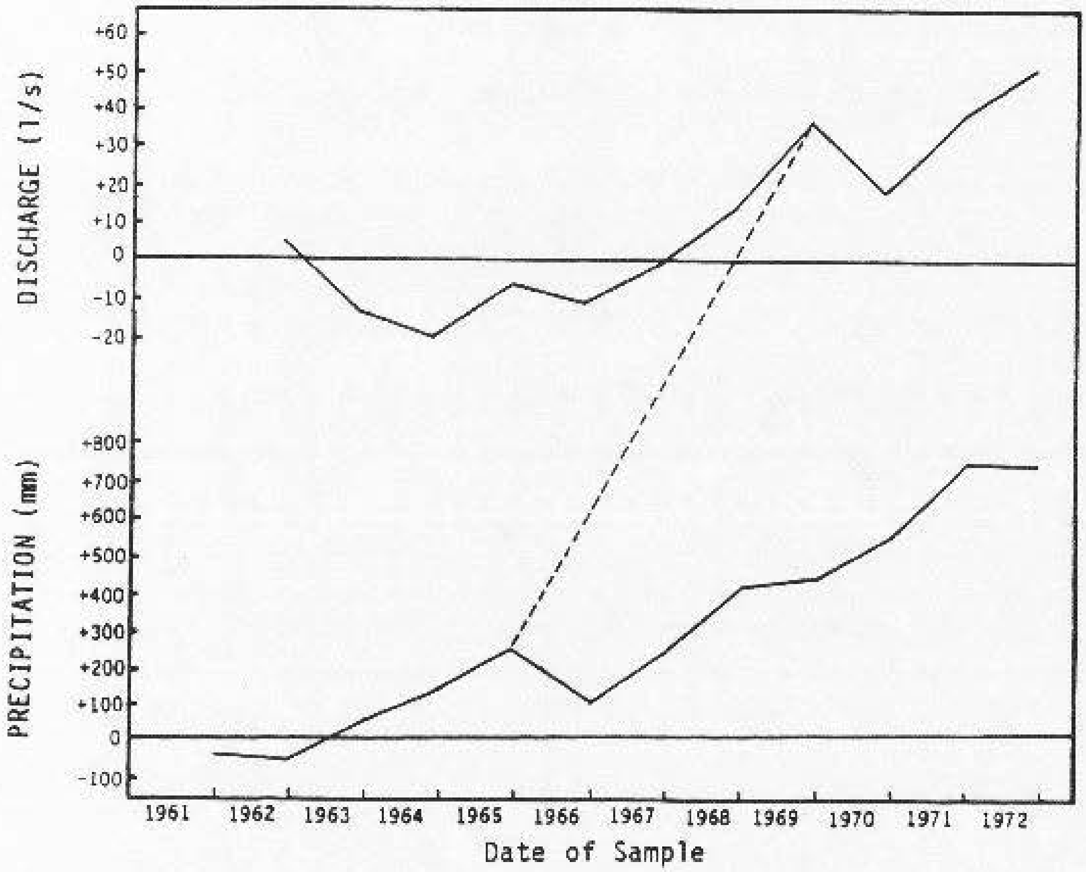


Figure 15.--Relation of annual departure from normal (1962-72) of Maple Spring discharge to cumulative annual departure from long-term average (1931-60) precipitation at Logan USU and Brigham City weather stations. Dotted line shows interpretation of correlation between precipitation and spring discharge.

(7 ft/day) since the tritium data suggest that the mean residence time is greater than six months. Quinlan and Ewers (1985) reported that flows in a mature carbonate aquifer, i.e., a well-developed karst, have velocities ranging from 9 to 2286 m/hr (30-7500 ft/hr). Groundwater velocities which range from 0.3 to 1.5 m/day (1-5 ft/day) represent flow through a terrane which is not a mature karst (Quinlan and Ewers, 1985). Figure 16 shows a conceptual model of Upper Maple Spring.

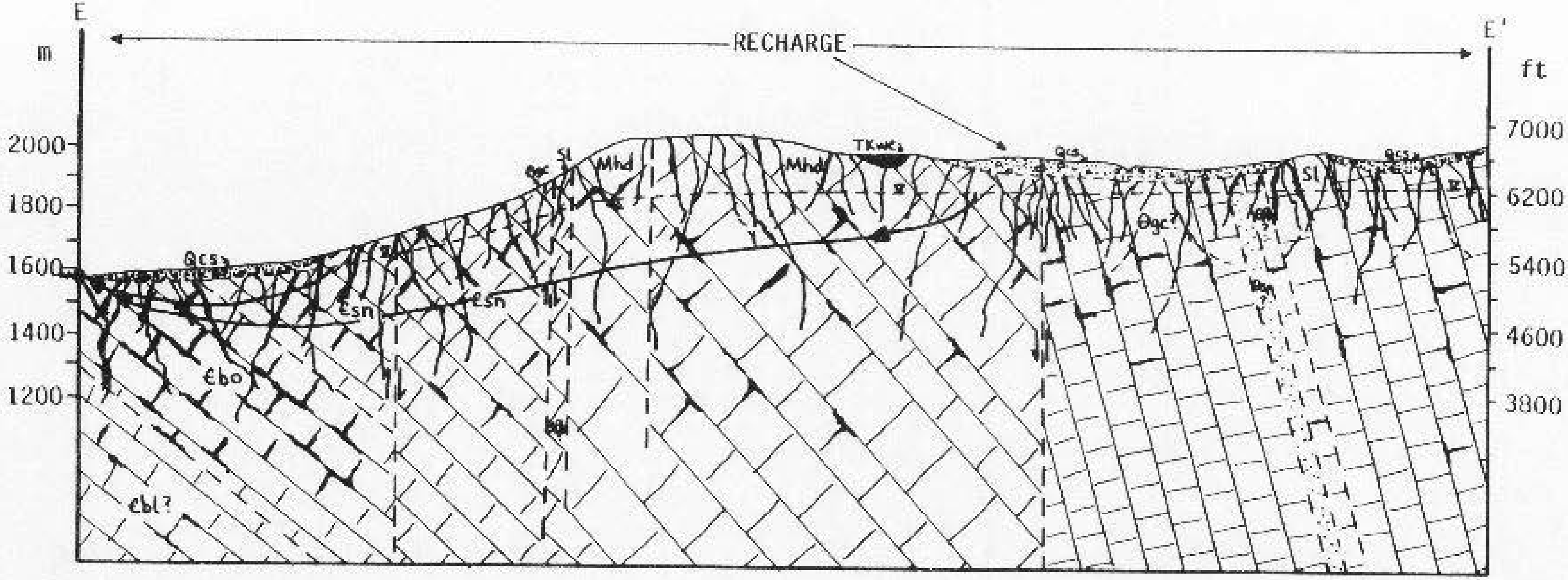


Figure 16.--Upper Maple Spring conceptual model. €bl, Blacksmith Dolostone; €bo, Bloomington Formation; €sn, St. Charles and Nounan Dolostones; Θgc, Garden City Formation; Θsp, Swan Peak Formation; Θfh, Fish Haven Dolostone; Sl, Laketown Dolostone; Mhd, Humbug and Deseret Formations; TKwe, Wasatch and Evanston Formations; Qcs, colluvium and slopewash, includes some alluvium; Geology adapted from Crittenden and Sorensen (1985). No vertical exaggeration.

CONCLUSIONS

A complex tectonic history of the study area has created down-dropped blocks which have formed valleys closed to surface-water drainage. The result of closed valleys may have been ponded water, which in turn created solution conduits for underground drainage in the predominantly carbonate terrane. Since valleys do not remain closed indefinitely to surface-water drainage, and recent climatic conditions have not been favorable for extensive karstification, karst features in the study area are rather poorly-developed.

Mantua Valley groundwater is generally a calcium-magnesium-bicarbonate water. Water sampled from five springs in Mantua Valley has low TDS and temperatures slightly above the mean annual air temperature of the area. Molar Ca^{+2}/Mg^{+2} ratios suggest that much of the groundwater supplying springs in Mantua Valley has cross-formational flow through dolostone and limestone formations.

Porosity and permeability of rocks in the study area are primarily the result of fracturing and solution along fractures and bedding planes. Aquifers in the study area are able to transmit relatively large amounts of groundwater because of a high fracture density and because joints and bedding planes have probably been widened by solution. Stratigraphic thicknesses of hydrostratigraphic units in the study area range from 3625 m to 4253 m for aquifers and from 2815 m to 3670 m for aquitards.

Peter Jensen Spring is probably controlled by a diffuse flow carbonate aquifer. Groundwater in the system probably circulates at depths greater than 140 m through the Upper Paleozoic carbonate aquifer. Recharge to the system is thought to be diffuse over approximately a 3.0 km^2 area to the east of the spring outlet.

West Hallings Spring appears to be controlled by an open-fracture system in parts of the Lower Paleozoic carbonate aquitard and parts of the Lower Paleozoic carbonate aquifer. Groundwater supplying West Hallings Spring seems to circulate at depths greater than 185 m and may have a mean residence time of approximately nine years. The recharge area for West Hallings Spring is approximately an 18.0 km^2 area to the northwest of the spring orifice, and recharge is thought to be predominantly diffuse.

Mud Spring also appears to be controlled by an open-fracture system through carbonate rocks but has a shallower depth of groundwater circulation (greater than 115 m) than West Hallings Spring. Groundwater supplying Mud Spring seems to be flowing through parts of the Lower, Middle, and Upper Paleozoic carbonate aquifers. The recharge area for Mud Spring is estimated to be a 9.0 km^2 area north and northeast of the spring orifice, and recharge is thought to be predominantly diffuse.

Groundwater flow to Mantua Spring seems to be controlled by a local flow system through narrow fractures in the Precambrian and Cambrian quartzite aquitard. Temperature data suggest that mixing of shallow and deeper groundwaters may occur along a zone of high permeability in the quartzites. Recharge to the system is probably diffuse, and the recharge area may be an approximately 4.0 km^2 area to the southwest of the spring outlet.

Maple Spring appears to be controlled by a complex open-fracture carbonate system. Groundwater supplying the spring probably flows through parts of the Lower, Middle, and Upper Paleozoic carbonate aquifers. Temperature data suggest that the depth of circulation of groundwater is relatively shallow (greater than 20 m). Tritium, calcite saturation indices, and precipitation-discharge data suggest that the mean residence time of Maple Spring groundwater may be approximately four years. The recharge area for Upper Maple Spring alone is estimated to be a 7.0 km^2 area (and 13.0 km^2 for Upper and Lower Maple Springs combined) to the southeast of the spring orifice. The system is thought to be recharged through a combination of diffuse and point-source recharge.

Because the conceptual models of the springs in this report are represented as two-dimensional models, it was necessary to greatly simplify the flow systems. However, the systems controlling flow to the springs are probably very complex because shallow flow components of the systems are controlled by rugged topography, the carbonate aquifers have probably experienced different degrees of modification by solution, mixing of shallow and deep groundwaters probably occurs, and some of the springs are probably supplied by diffuse as well as point-source recharge.

SUGGESTIONS FOR FURTHER STUDY

Chemical analyses of monthly samples from Dry Lake as well as springs to the east and north of the valley might determine which springs in the area are recharged by Dry Lake Valley. If Dry Lake water became highly saline at low stands, a geochemical signature of the lake water would be created. If the residence times of the systems are short enough, it is possible that Dry Lake water could be traced to one or more of the springs in the area through the chemistries of the waters.

To obtain a better understanding of the residence times of groundwater in the flow systems, frequent sampling (monthly, for at least six months), and analysis for environmental tritium is recommended. Once 1984-86 tritium data for the Salt Lake City, Utah station are published, they may alter or confirm the interpretation of the residence times of West Hallings and Maple Springs.

If snow was sampled at various elevations in the study area and analyzed for $^{18}O/^{16}O$ ratios, and if groundwater from the springs was sampled and analyzed for $^{18}O/^{16}O$ ratios, it might be possible to estimate the elevations at which recharge to each of the springs occurs.

A statistical correlation of Mud Spring discharge (available from Bjorklund and McGreevy, 1974, fig. 6, p. 23) and precipitation data might define the relationship between recharge events and Mud Spring discharge rates. Discharge measurements of Mud and Upper Maple Springs might be possible to make, since they discharge at their natural outlets. By using TDS and discharge data for the springs, a mixing model (Hall, 1970) could be developed for each spring, allowing a better understanding of their complex flow systems.

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APPENDICES

Appendix A. Mantua Valley Groundwater Chemistry Analyses

Station	Date Sampled	T °C	pH	Na^+ mg/l	K^+ mg/l	Ca^{+2} mg/l	Mg^{+2} mg/l	HCO_3^- mg/l	SO_4^{-2} mg/l	Cl^- mg/l	SiO_2 mg/l	TDS mg/l
Peter Jensen	3-21-85	9.8	6.5	7.5	0.8	56.1	18.9	n.d.	n.d.	10	n.d.	n.d.
Peter Jensen	5-03-85	12.5	6.7	7.2	0.9	50.3	19.1	n.d.	9	8	7.8	262
Peter Jensen	6-07-85	12.6	7.0	6.9	0.7	50.6	19.5	n.d.	8	9	8.5	250
Peter Jensen	7-17-85	12.8	6.9	6.7	0.7	51.8	19.8	234	9	5	8.2	264
Peter Jensen	9-26-85	11.5	7.2	8.2	0.7	54.5	17.9	242	9	8	8.9	254
Peter Jensen	11-24-85	10.2	7.5	7.3	0.7	53.9	17.9	267	11	8	8.1	268
Peter Jensen	1-19-86	11.0	7.1	7.3	0.7	50.1	17.8	284	10	8	11	248
Peter Jensen	3-15-86	10.7	7.0	7.9	0.8	51.5	19.0	250	10	8	9.5	250
Peter Jensen	5-25-86	11.6	7.0	7.1	0.8	59.4	21.3	284	9	8	8.2	252
West Hallings	3-21-85	11.5	6.6	10.0	1.1	48.3	21.0	n.d.	n.d.	15	n.d.	n.d.
West Hallings	5-03-85	14.4	6.9	8.0	1.8	40.2	20.2	n.d.	8	10	11	256
West Hallings	6-07-85	18.9	7.1	7.9	0.8	40.8	20.5	n.d.	8	11	10	244
West Hallings	7-17-85	13.9	7.3	7.5	1.2	40.0	19.8	250	8	7	11	258
West Hallings	9-26-85	14.1	7.1	9.2	1.1	46.5	18.9	217	8	11	11	246
West Hallings	11-24-85	13.0	7.6	8.6	1.0	45.9	18.2	250	10	11	9.6	244
West Hallings	1-19-86	14.1	7.2	8.8	1.1	42.3	17.9	250	7	11	13	240
West Hallings	3-15-86	12.3	7.0	8.5	1.1	44.0	19.4	242	8	11	11	239
West Hallings	5-25-86	13.4	7.5	8.2	1.0	51.1	22.5	259	8	10	11	250
Mud	3-21-85	9.4	6.9	6.7	0.7	54.2	22.2	n.d.	n.d.	9	n.d.	n.d.
Mud	7-14-85	12.7	7.5	6.1	1.6	45.6	21.2	250	9	6	9.3	264
Mud	9-24-85	12.1	7.1	6.5	0.7	49.4	18.9	242	9	7	9.1	248
Mud	11-24-85	10.7	7.7	6.2	1.1	49.1	19.6	225	10	8	8.4	252
Mud	1-19-86	11.0	7.8	7.3	1.2	46.6	19.5	250	8	8	11	244
Mud	3-15-86	10.8	6.9	7.1	0.8	45.9	20.3	259	9	8	9.3	256
Mud	5-25-86	12.9	7.0	7.3	0.9	54.1	23.3	250	8	7	9.2	244
Mantua	4-02-85	5.6	5.9	4.5	0.7	7.7	2.0	n.d.	n.d.	4	n.d.	n.d.
Mantua	5-03-85	8.7	6.0	4.0	1.5	6.7	2.0	n.d.	6	2	7.8	80
Mantua	7-17-85	11.2	6.5	3.6	0.7	7.3	1.8	42	6	2	7.3	66
Mantua	9-24-85	12.7	6.5	3.9	0.5	7.2	2.1	42	6	3	7.1	56
Mantua	11-24-85	5.2	7.4	5.7	0.7	12.5	4.2	75	9	5	6.9	82
Mantua	1-19-86	7.0	6.8	4.8	0.7	13.5	6.0	92	5	4	9.6	96
Mantua	3-15-86	7.9	7.0	7.4	1.1	24.9	14.5	175	7	7	12	168
Mantua	5-25-86	7.8	5.7	3.8	0.8	6.3	1.9	33	4	2	7.4	58
Maple	3-21-85	6.6	6.8	5.8	0.8	43.1	15.9	n.d.	n.d.	9	n.d.	n.d.
Maple	5-03-85	9.9	7.4	5.6	1.4	31.6	13.4	n.d.	6	5	8.0	178
Maple	6-07-85	10.1	8.4	5.9	1.1	35.5	14.5	n.d.	6	7	8.6	198
Maple	7-14-85	8.0	7.5	5.7	0.7	37.0	15.6	242	5	5	8.3	210
Maple	9-24-85	8.8	7.1	6.9	0.7	42.4	15.1	217	8	6	8.4	196
Maple	11-24-85	7.7	7.6	6.7	1.2	42.0	14.4	209	9	6	7.6	194
Maple	1-19-86	8.0	7.4	6.7	0.9	39.1	14.2	225	6	7	10	198
Maple	3-15-86	7.3	6.8	8.1	1.0	35.7	11.9	200	6	6	8.5	176
Maple	5-25-86	9.1	7.3	7.1	0.7	45.0	14.5	209	6	5	7.9	196
Well	3-21-85	n.d.	n.d.	13.3	0.9	63.3	22.7	n.d.	n.d.	14	n.d.	n.d.
Well	5-03-85	12.8	6.7	13.3	2.3	53.3	24.8	n.d.	9	11	13	308
Well	6-07-85	16.3	7.2	14.6	0.7	53.3	23.7	n.d.	8	13	13	313
Well	7-14-85	13.7	6.9	13.6	1.6	54.2	24.1	334	8	18	12	322
Well	9-24-85	14.6	7.1	14.4	0.9	58.7	19.3	309	7	8	15	294
Well	11-24-85	6.8	7.8	13.3	1.0	58.3	20.3	301	10	10	12	304
Well	1-19-86	6.4	7.7	13.4	0.9	54.7	21.4	342	8	12	15	304
Well	3-15-86	6.5	7.1	15.6	0.9	55.1	22.0	294	9	11	14	300
Well	5-25-86	12.9	7.6	14.7	1.0	65.0	25.7	317	8	11	14	324

Appendix B. Data for Secondary Porosity Estimates in Outcrops

Formation	Location in Study Area	Average width of fractures (cm)	Number of fractures intersected	Length of Sampling Line (m)
Cambrian Geertson Canyon Quartzite	SE 1/4 of SE 1/4 of sec. 21, T. 9 N., R. 1 W.	0.5 0.75	11 16	0.9 1.5
Cambrian Langston Dolostone	NW 1/4 of NW 1/4 of sec. 16, T. 9 N., R. 1 W.	0.25	8	0.6
Cambrian Blacksmith Dolostone	NW 1/4 of SW 1/4 of sec. 9, T. 9 N., R. 1 W.	0.25 1.3	60 12	1.5 1.5
Cambrian Bloomington Formation	SW 1/4 of SE 1/4 of sec. 6, T. 8 N., R. 1 W.	7.5	7	2.3
Cambrain St. Charles Dolostone	SE 1/4 of NE 1/4 of sec. 5, T. 9 N., R. 1 W.	5.0	2	0.9
Ordovician Garden City Formation	NW 1/4 of NW 1/4 of sec. 12, T. 8 N., R. 1 W.	5.0	6	1.8
Ordovician Swan Peak Formation	SW 1/4 of SE 1/4 of sec. 10, T. 9 N., R. 1 W.	0.75 0.75	24 7-18	9.5 4.0
Silurian Laketown Dolostone	SE 1/4 of SE 1/4 of sec. 1, T. 8 N., R. 1 W.	7.5 17.7	8 4	4.9 4.9
Mississippian Lodgepole Limestone	SW 1/4 of SE 1/4 of sec. 3, T. 9 N., R. 1 W.	2.5	6	1.8
Mississippian Humbug/Deseret Formations, undivided	SE 1/4 of NE 1/4 of sec. 1, T. 8 N., R. 1 W.	2.5 4.6	10 9	2.3 3.0
Mississippian Brazer Formation	SE 1/4 of SE 1/4 of sec. 28, T. 10 N., R. 1 W.	1.5 2.0	16 11	2.2 2.2

Appendix C. Saturation Indices of Groundwater Samples with Respect to Calcite, Dolomite, and Quartz

Station	Date Sampled	Saturation Index		
		Calcite	Dolomite	Quartz
Peter Jensen	7-17-85	-0.536	-1.317	0.335
Peter Jensen	9-26-85	-0.321	-0.977	0.392
Peter Jensen	11-24-85	**-0.009**[a]	-0.371	0.373
Peter Jensen	1-19-86	-0.399	-1.107	0.656
Peter Jensen	3-15-86	-0.543	-1.386	0.434
Peter Jensen	5-25-86	-0.424	-1.142	0.355
West Hallings	7-17-85	-0.299	-0.712	0.444
West Hallings	9-26-85	-0.491	-1.178	0.441
West Hallings	11-24-85	**0.041**	-0.144	0.399
West Hallings	1-19-86	-0.371	-0.921	0.514
West Hallings	3-15-86	-0.596	-1.385	0.471
West Hallings	5-25-86	**0.002**	-0.169	0.452
Mud	7-14-85	**-0.067**	-0.295	0.391
Mud	9-24-85	-0.452	-1.161	0.392
Mud	11-24-85	**0.089**	**-0.085**	0.380
Mud	1-19-86	**0.217**	**0.193**	0.492
Mud	3-15-86	-0.675	-1.569	0.423
Mud	5-25-86	-0.494	-1.180	0.384
Mantua	7-17-85	-2.472	-5.409	0.311
Mantua	9-24-85	-2.529	-5.423	0.274
Mantua	11-24-85	-1.282	-3.015	0.389
Mantua	1-19-86	-1.737	-3.764	0.502
Mantua	3-15-86	-1.021	-2.197	0.583
Mantua	5-25-86	-3.561	-7.564	0.375
Maple	7-14-85	-0.228	-0.748	0.421
Maple	9-24-85	-0.603	-1.557	0.413
Maple	11-24-85	-0.142	-0.672	0.388
Maple	1-19-86	-0.333	-1.024	0.502
Maple	3-15-86	-1.024	-2.458	0.444
Maple	5-25-86	-0.388	-1.164	0.381
Well	7-14-85	-0.470	-1.104	0.486
Well	9-24-85	-0.249	-0.779	0.568
Well	11-24-85	**0.310**	**0.221**	0.602
Well	1-19-86	**0.229**	**0.102**	0.706
Well	3-15-86	-0.437	-1.219	0.675
Well	5-25-86	**0.265**	**0.302**	0.565

[a]Saturated and supersaturated values with respect to calcite and dolomite are in bold print

Appendix D. Tritium and Adjusted Tritium Data, 1963-1983,

Salt Lake City, Utah Station

Sample Date	Activity of Sample on Sample Date (TU)	Precipitation at Salt Lake City station (mm)	Activity of Sample, weighted[a] for precipitation (TU)	Age of Water, 6-86 (yrs)	Activity, 6-86, of weighted sample, adjusted for radioactive decay (TU)
1963	3577.5	--	--	23	992.2[b]
1964	2561.8	--	--	22	751.3
1965	965.7	--	--	21	299.4
1966	599.2	--	--	20	196.4
1967	463.9	--	--	19	160.8
1968	312.0	--	--	18	114.4
1969	268.5	--	--	17	104.1
1-70	71.8	32	54.4	16.4	21.8
2-70	143.0	24	81.2	16.3	32.7
3-70	261.0	26	160.6	16.25	64.9
4-70	346.0	83	679.7	16.2	275.4
5-70	430.0	23	234.1	16.1	95.4
6-70	459.0	41	445.4	16.0	182.5
7-70	229.0	22	119.2	15.9	49.1
8-70	190.0	15	67.5	15.8	28.0
9-70	195.0	71	327.7	15.75	136.2
10-70	114.0	41	110.6	15.7	46.1
11-70	92.3	58	126.7	15.6	53.1
12-70	n.d.	71	n.d.	15.5	n.d.
1-71	77.0	27	52.3	15.4	22.2
2-71	110.0	54	149.4	15.3	63.7
3-71	237.0	26	155.0	15.25	66.2
4-71	440.0	55	608.8	15.2	260.8
5-71	459.0	34	392.6	15.1	169.2
6-71	447.0	16	179.9	15.0	77.9
7-71	213.0	24	128.6	14.9	56.0
8-71	164.0	55	226.9	14.8	99.4
9-71	155.0	44	171.6	14.75	75.4
10-71	95.5	82	197.0	14.7	86.8
11-71	96.9	26	63.4	14.6	28.1
12-71	91.6	34	78.3	14.5	34.9
1-72	54.2	31	50.2	14.4	22.5
2-72	104.0	12	37.3	14.3	16.8
3-72	120.0	30	107.5	14.25	48.6
4-72	128.0	92	351.5	14.2	159.2

Sample Date	Activity of Sample on Sample Date (TU)	Precipitation at Salt Lake City station (mm)	Activity of Sample, weighted for precipitation (TU)	Age of Water, 6-86 (yrs)	Activity, 6-86, of weighted sample, adjusted for radioactive decay (TU)
5-72	164.0	4	19.6	14.1	8.9
6-72	n.d.	4	n.d.	14.0	n.d.
7-72	153.0	2	9.1	13.9	4.2
8-72	95.9	5	14.3	13.8	6.6
9-72	69.4	35	72.5	13.75	33.7
10-72	50.9	70	106.4	13.7	49.6
11-72	55.7	35	58.2	13.6	27.3
12-72	61.9	82	151.5	13.5	71.4
1-73	53.7	38	47.3	13.4	22.4
2-73	65.3	23	34.8	13.3	16.6
3-73	104.0	68	164.1	13.25	78.4
4-73	105.0	42	102.3	13.2	49.0
5-73	123.0	44	125.6	13.1	60.5
6-73	139.0	5	16.1	13.0	7.8
7-73	125.0	27	78.3	12.9	38.1
8-73	92.4	29	62.2	12.8	30.5
9-73	78.3	103	187.1	12.75	91.9
10-73	62.4	17	24.6	12.7	12.1
11-73	36.2	64	53.8	12.6	26.6
12-73	46.0	57	60.8	12.5	30.3
1-74	45.8	46	68.5	12.4	34.3
2-74	65.3	42	89.2	12.3	44.9
3-74	81.3	25	66.1	12.25	33.4
4-74	150.0	116	565.9	12.2	286.6
5-74	327.0	10	106.3	12.1	54.1
6-74	162.0	7	36.9	12.0	18.9
7-74	81.5	5	13.3	11.9	6.8
8-74	n.d.	8	n.d.	11.8	n.d.
9-74	n.d.	1	n.d.	11.75	n.d.
10-74	58.9	52	99.6	11.7	51.9
11-74	48.3	23	36.1	11.6	18.9
12-74	45.1	34	49.9	11.5	26.3
1-75	54.9	33	47.8	11.4	25.3
2-75	84.4	31	69.0	11.3	36.7
3-75	85.4	87	196.0	11.25	104.7
4-75	117.0	62	191.4	11.2	102.5
5-75	136.0	66	236.8	11.1	127.5
6-75	143.0	46	173.6	11.0	94.0
7-75	73.1	7	13.5	10.9	7.4

Sample Date	Activity of Sample on Sample Date (TU)	Precipitation at Salt Lake City station (mm)	Activity of Sample, weighted for precipitation (TU)	Age of Water, 6-86 (yrs)	Activity, 6-86, of weighted sample, adjusted for radioactive decay (TU)
8-75	20.7	3	1.6	10.8	0.9
9-75	39.8	2	2.1	10.75	1.2
10-75	35.9	49	46.4	10.7	25.6
11-75	35.7	43	40.5	10.6	22.4
12-75	36.6	26	25.1	10.5	14.0
1-76	35.9	16	22.1	10.4	12.4
2-76	39.1	48	72.2	10.3	40.7
3-76	54.3	48	100.2	10.25	56.6
4-76	69.9	63	169.4	10.2	95.9
5-76	103.0	25	99.0	10.1	56.4
6-76	n.d.	31	n.d.	10.0	n.d.
7-76	49.9	39	74.9	9.9	43.1
8-76	50.7	21	41.0	9.8	23.7
9-76	36.6	4	5.6	9.75	3.3
10-76	n.d.	14	n.d.	9.7	n.d.
11-76	n.d.	1	n.d.	9.6	n.d.
12-76	37.1	2	2.9	9.5	1.7
1-77	41.2	19	21.0	9.4	12.4
2-77	35.5	16	15.2	9.3	9.0
3-77	59.6	79	126.2	9.25	75.3
4-77	80.4	15	32.3	9.2	19.3
5-77	127.0	121	412.0	9.1	248.0
6-77	n.d.	2	n.d.	9.0	n.d.
7-77	67.0	15	26.9	8.9	16.4
8-77	59.3	47	74.7	8.8	45.7
9-77	72.6	47	91.5	8.75	56.2
10-77	40.9	21	23.0	8.7	14.2
11-77	45.0	30	36.2	8.6	22.4
12-77	46.5	36	44.9	8.5	28.0
1-78	54.8	59	84.2	8.4	52.7
2-78	91.1	50	118.6	8.3	74.7
3-78	97.2	88	222.8	8.25	140.6
4-78	103.0	74	198.5	8.2	125.7
5-78	229.0	40	238.5	8.1	151.8
6-78	n.d.	2	n.d.	8.0	n.d.
7-78	n.d.	2	n.d.	7.9	n.d.
8-78	70.6	23	42.3	7.8	27.4
9-78	52.1	64	86.8	7.75	56.3
10-78	n.d.	0	n.d.	7.7	n.d.

Sample Date	Activity of Sample on Sample Date (TU)	Precipitation at Salt Lake City station (mm)	Activity of Sample, weighted for precipitation (TU)	Age of Water, 6-86 (yrs)	Activity, 6-86, of weighted sample, adjusted for radioactive decay (TU)
11-78	n.d.	44	n.d.	7.6	n.d.
12-78	n.d.	15	n.d.	7.5	n.d.
1-79	n.d.	18	n.d.	7.4	n.d.
2-79	n.d.	27	n.d.	7.3	n.d.
3-79	40.4	20	44.2	7.25	29.5
4-79	n.d.	26	n.d.	7.2	n.d.
5-79	n.d.	21	n.d.	7.1	n.d.
6-79	n.d.	9	n.d.	7.0	n.d.
7-79	n.d.	10	n.d.	6.9	n.d.
8-79	n.d.	16	n.d.	6.8	n.d.
9-79	n.d.	1	n.d.	6.75	n.d.
10-79	n.d.	33	n.d.	6.7	n.d.
11-79	n.d.	25	n.d.	6.6	n.d.
12-79	n.d.	14	n.d.	6.5	n.d.
1-80	22.9	73	45.9	6.4	32.1
2-80	12.5	57	19.6	6.3	13.8
3-80	13.7	62	23.3	6.25	16.4
4-80	33.4	23	21.1	6.2	14.9
5-80	71.9	69	136.3	6.1	97.0
6-80	n.d.	11	n.d.	6.0	n.d.
7-80	29.8	34	27.8	5.9	20.0
8-80	33.7	7	6.5	5.8	4.7
9-80	21.1	18	10.4	5.75	7.5
10-80	32.2	44	38.9	5.7	28.3
11-80	18.6	30	15.3	5.6	11.2
12-80	n.d.	9	n.d.	5.5	n.d.
1-81	n.d.	16	n.d.	5.4	n.d.
2-81	54.9	21	32.9	5.3	24.5
3-81	65.1	54	100.4	5.25	74.9
4-81	88.9	11	27.9	5.2	20.9
5-81	79.0	93	210.0	5.1	158.0
6-81	n.d.	26	n.d.	5.0	n.d.
7-81	n.d.	8	n.d.	4.9	n.d.
8-81	53.0	6	9.1	4.8	7.0
9-81	27.2	12	9.3	4.75	7.2
10-81	18.9	99	53.5	4.7	41.2
11-81	19.7	26	14.6	4.6	11.3
12-81	17.2	48	23.6	4.5	18.4

Sample Date	Activity of Sample on Sample Date (TU)	Precipitation at Salt Lake City station (mm)	Activity of Sample, weighted for precipitation (TU)	Age of Water, 6-86 (yrs)	Activity, 6-86, of weighted sample, adjusted for radioactive decay (TU)
1-82	20.7	27	11.6	4.4	9.1
2-82	28.5	13	7.7	4.3	6.1
3-82	24.5	61	31.0	4.25	24.5
4-82	26.1	41	22.2	4.2	17.6
5-82	33.7	47	32.8	4.1	26.1
6-82	49.0	17	17.3	4.0	13.8
7-82	23.4	65	31.5	3.9	25.3
8-82	17.8	14	5.2	3.8	4.2
9-82	16.8	179	62.3	3.75	50.5
10-82	16.6	47	16.2	3.7	13.2
11-82	15.3	19	6.0	3.6	4.9
12-82	15.3	49	15.5	3.5	12.8
1-83	15.7	30	9.2	3.4	7.6
2-83	14.8	35	10.1	3.3	8.4
3-83	19.2	101	37.7	3.25	31.5
4-83	24.5	41	19.5	3.2	16.3
5-83	32.4	66	41.6	3.1	35.0
6-83	33.3	16	10.4	3.0	8.8
7-83	17.1	26	8.6	2.9	7.3
8-83	13.1	67	17.1	2.8	14.6
9-83	30.8	26	15.6	2.75	13.4
10-83	20.1	41	16.0	2.7	13.8
11-83	11.2	57	12.4	2.6	10.7
12-83	11.3	111	24.4	2.5	21.2

[a]weighted by:

$$\frac{\text{activity of sample X monthly precipitation}}{\text{monthly mean of the year's precipitation}} = \text{activity of sample weighted for precipitation}$$

(see discussion in text, p. 30)

[b]1963-1969 data are adjusted for radioactive decay but not weighted for precipitation